The mass balance modelling of Spitsbergen glaciers

by

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This thesis is submitted as a partial requirement for the degree of Master of Philosophy in Polar Studies of the University of Cambridge.

June 1992

Fitzwilliam College

Contents

Declaration

I declare that no part of this thesis has been accepted or presented for the award of any degree or diploma by any university, and that to the best of my knowledge, this thesis contains no material previously published or written by any other person or persons, except where due reference in given to the author or authors by direct credit in the text or bibliography.

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Abstract

This thesis deals with a series of mass balance studies carried out on two glaciers in north-west Spitsbergen, Midre Lovénbreen and Austre Brøggerbreen, using an energy balance model. A general description of glaciers, including their classification and the manner in which their surfaces may be sub-divided, is carried out. A discussion is then made concerning the processes which affect the mass balance of a glacier surface, and the techniques used, both field and computational, to determine this parameter.

The first modelling exercises carried out were to determine how sensitive the model used is to perturbations in various climatic parameters. This project had a number of objectives: 1) determining what difference is made by applying simplified meteorological data instead of measured data, 2) modelling the mass balances of Midre Lovénbreen and Austre Brøggerbreen to reproduce the values obtained by field surveys, 3) modelling the mass balances of these glaciers to determine the effect changes representing future warmer, and past cooler, climates would have on their balance behaviour.

It was found that temperature was the parameter the model results were most sensitive to, and cloud height the least. There is a difference in the final results depending upon what meteorological data form is used, simplified or measured. The mass balances of the glaciers could be best reproduced by a model version that included an altitudinal precipitation correction expression and excluded the latent heat component of the energy equation. A rise in temperature of 4 to 5°C will result in most of the glacier's present extent becoming areas of negative balance, despite applying a 50% increase in precipitation, while a decrease in temperature of 2 to 3°C will considerably increase the area which will experience a positive balance, sometimes to lower than the glaciers' present lowest point.

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Acknowledgments

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Thanks to Miss Anna Jolly who translated one of my references from German.

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Chapter One
Introduction

1.1 Glaciers, climate and sea level

The interest in both public and scientific circles regarding the consequences of global warming, whether by greenhouse gases or natural cycles, provides the motivation to pursue the study of glaciers and their response to climatic variation.

This motivation results from two factors. The first is the fact that glaciers provide a great deal of information with regards to the identification of changes in a region's climate. Glaciers are very sensitive to climatic change (Oerlemans, 1988; Koster, 1991), especially when these changes alter the energy balance of their surfaces. Examination of changes in the position of glacier fronts provides evidence for climatic anomalies and this has inspired several programs which deal with the observation of glacier behaviour around the world (Dyurgerov, 1990). There has been a world wide trend observed for the past 100 to 150 years for a decrease in the extent of glaciers, since the end of the so-called Little Ice Age. This is apparent not only from recent studies of mass balance, but from observations using photographic and mapping exercises.

The second factor concerns the contribution to the expected rise in sea level by the melting of glaciers, excluding those associated with the Antarctic or Greenland ice sheets. The global sea level appears to have risen over the past 100 years by between 10 to 15 cm (Meier, 1984). Part of this is due to the thermal expansion of the oceans resulting from the accompanying rise in the global mean temperature, thought to be between 0.3 - 0.6°C over the past century (Koster, 1991). The remaining sea level rise is considered to result from the melting of ice masses and changes in the amount of water stored on land in surface reservoirs and aquifers. However, the mass balance conditions of the Antarctic and Greenland ice sheets, the largest reservoirs of ice, are uncertain. They may be balanced or even experiencing a positive balance (Meier, 1990).

Some estimates of the various contributors to sea level variation are; the growth of the Greenland and Antarctic ice sheets resulting in a sea level decrease of 0.45 +/- 0.25 mm yr^{-1} and 0.75 +/- 0.25 mm yr^{-1} respectively, thermal expansion of the oceans accounting for an increase of 0.2 +/- 0.1 mm yr^{-1}, changes in the land liquid water content account for less than 0.2 mm yr^{-1} and increases due to the melting of land ice by 0.5 +/- 0.3 mm yr^{-1}. It is obvious from these values, and the associated error, that the processes involved are still not fully understood (Meier, 1990). In fact, it is the lack of reliable data which is the main difficulty when resolving the problem of the extent of sea level change, and its possible causes (Meier, 1984; Meier, 1990).

Any melting which would occur because of the increased temperatures may be off-set by increased precipitation over the ice masses. There would also be a delay in the expected rise in sea level because the rate at which the resulting melt water would run off into the ocean would be slowed by its percolating and refreezing in the snowpack. This delay may last for decades in the Arctic and Antarctic ice masses (Meier, 1990).

1.2 Objectives and study area of this project

1.2.1 Objectives of the project

The purpose of this project is to model the mass balance behaviour of the surface of Spitsbergen glaciers. This is done using a surface energy balance model developed by Professor Johannes Oerlemans of the Institute for Oceanography and Meteorology, University of Utrecht, the Netherlands. It is a version of a computer program utilised in Oerlemans (1991) which was used to calculate the mass balance of different regions of the Greenland Ice Sheet.

The objectives of this project were as follows:

1. Oerlemans (pers. comm. and 1991) made a number of assumptions with regards to the climatic parameters in his model. Amongst these were that precipitation, humidity and cloudiness are constant throughout the balance year. Humidity and cloudiness are defined to be the annual average, while it was assumed that there was precipitation on every

second day, the amount of which was again constant. Daily temperature is calculated for the model using a sinusoidal function, with the annual mean and range needing to be defined. A series of tests were carried out to determine if the mass balance results obtained using these assumptions differed from those gained when measured meteorological data was used in the model.

2. The mass balances of two Spitsbergen glaciers were modelled. These exercises made use of recorded meteorological data from a weather station located close to the glaciers. The results were compared with values measured by glaciological surveys conducted by members of the Norsk Polarinstitutt (Liestøl, 1986; Hagen and Liestøl, 1987; Hagen, 1988).

3. The next objective was to determine what effect different climatic scenarios would have on the mass balance of Spitsbergen glaciers. These scenarios were the effect of greenhouse gas induced warming, and the cooler conditions of the Little Ice Age. The climatic parameters varied for the warmer conditions were the annual mean temperature and precipitation, while annual mean temperature was varied for the cooler situation. There were also tests carried out to determine the effect of a change in the intra-annual climatic variability of the area. By variability, it is meant how climatic parameters such as temperature vary on a daily basis. This variability is especially noticeable when examining the average daily temperatures over a year (appendix Two).

1.2.2 The project study area

The computer model used in this project was applied to two glaciers located in the north-west of Spitsbergen, part of the Svalbard archipelago (figure 1.1). The two glaciers were Austre Brøggerbreen and Midre Lovénbreen. They are small cirque glaciers with areas of approximately 6 km^2. The reason for choosing these particular ice masses is that they have been extensively studied by members of the Norsk Polarinstitutt, and are located close (within 3 to 5 km) of the Norsk Polarinstitutt weather station at Ny-Ålesund (Norsk Polarinstitutt, 1979). The surveys conducted on these glaciers include mass balance measurements, seismic, gravity, ground and airborne radio echo soundings. There is also a continuous time series of climatic data from the Ny-Ålesund weather station since mass balance

Figure 1.1 A map showing the Svalbard archipelago (from Hisdal, 1985). The insert is the location of the glaciers studied during this project, Austre Brøggerbreen and Midre Lovénbreen, in the Kongsfjorden area.

measurements were started in the late 1960's. Five balance years of weather data from this station, 1983 to 1988, were used during this project.

The model used in this project could in the future be incorporated into a more comprehensive dynamic ice mass model. Numerical models of glaciers depend upon the surface mass balance regardless of their complexity (Oerlemans, 1991) and any change in the climate of a glacier's area would be reflected in this parameter. This proposed future model would take into account the time dependent flow of the ice and the various feed-back effects which are still a topic of uncertainty. It in turn may be part of a larger model which incorporates coupling between the glacier and atmospheric processes.

1.3 Outline of the thesis

This chapter serves to introduce the project and the glaciological problem to be treated, together with some of the reasons for pursuing it. Chapter Two will deal with the mass balance of glaciers. Some schemes used to classify glaciers will be discussed, followed by a description of the various snow and ice zones or facies into which a glacier may be divided. The main factors and processes which affect the mass and energy balance of a glacier will then be described.

Chapter Three is concerned with determining the mass balance of a glacier's surface. Field techniques will be described, as will some indirect ways of assessing a glacier's mass balance by the use of other parameters. These include the equilibrium line altitude and statistical correlation between measured mass balances and meteorological data from nearby weather stations. The model of Oerlemans (1991), which was used during this project, will be described, including its formulation and the assumptions made in its operation.

Chapter Four contains background about the island of Spitsbergen and the glaciers of Midre Lovénbreen and Austre Brøggerbreen (figure 1.1) which are modelled during this project. Topics discussed will include the topography, climate and glacial environment of Spitsbergen. The weather station at Ny-Ålesund will be mentioned, as it is from here that the climatic data used were acquired. Some details of the glaciers, and the work which has been undertaken on them by the Norsk Polarinstitutt, will follow.

Chapters Five and Six are concerned with the modelling procedures used and their results. Chapter Five will describe the procedures carried out. These include tests made to determine what parameters the model is most sensitive to, and how the assumptions of Oerlemans (1991) regarding the climatic data affected the results. Exercises were next carried out to reproduce the mass balances recorded by field measurements of the two glaciers studied. There were also tests to determine what effect greenhouse warming and the cool conditions of the Little Ice Age would have on the mass balance of these glaciers in comparison with the present day situation. The results and interpretations are presented in Chapter Six.

Chapter Seven is the conclusion, with a summary of the results found and some recommendations for future work.

There are several appendices which include the model program itself, meteorological data from Ny-Ålesund and details of the results.

Chapter Two
Glaciers and their mass balance

This chapter deals with the mass balance of glaciers. First, there is a summary of how glaciers may be classified. This is followed by a description of the various zones into which a glacier's surface may be divided. The mass and energy balances of glacier surfaces will then be discussed. This will cover the accumulation and mass loss processes and the various components of the energy budget of a glacier surface.

2.1 Classification of glaciers

Glaciers may be classified according to their relationship with the surrounding topography. *Ice sheets* and *ice caps* are imposed upon a region's topography, where the direction and rate of ice flow is defined by the ice mass's shape rather than the topography. *Ice shelves* are floating masses constrained by a coastline, while glaciers such as those studied during this project have their shape and flow constrained by the topography (Sugden and John, 1976).

Alhmann (1935) classified glaciers according to the temperature of the glacier ice and the amount of surface melting which occurs. This grouping is *temperate* and *polar*. Temperate glaciers are at the pressure melting point temperature throughout their mass, except for the upper few metres during winter.

Polar glaciers are divided further into *sub-polar* and *high-polar*. These glaciers are made up of firn extending, for sub-polar glaciers to 10 or 20 metres below the glacier's surface, and for high-polar glaciers sometimes to deeper than 100 metres (Alhmann, 1935). In high-polar glaciers, there is no surface melting, not even in firn, while there is some on sub-polar glacial surfaces. The term *cold* is used to classify polar glaciers by Paterson (1981). It is also noted in Paterson (1981) that temperate is not a precise term, and that these classifications should be restricted to specific areas of a glacier's mass, and that no attempt should be made to place an entire glacier into a single category.

Kuhn (1984) offers a climatic classification scheme for glaciers based upon the differences in their mass balance profiles (mass balance variation with altitude) for positive and negative mass balance years, that is, years where the glacier is increasing or decreasing in bulk. Alpine glaciers show differences between their positive and negative mass balance profiles which are almost independent of altitude. The altitude dependent differences which do occur are between the respective equilibrium line altitudes where the surface albedo decreases going from positive to negative balance years. Positive and negative balance differences for continental glaciers are again most obvious around the equilibrium line altitude where the change in albedo is the greatest. Maritime glaciers show their differences mainly in the accumulation area, the influence of increased solid precipitation in positive years being the dominant factor. Polar glaciers show changes mainly around the ablation area where the variation of the ablation season from positive to negative years is the most significant influence.

2.2 Zones of a glacier

Figure 2.1 shows the various sections into which the accumulation zone of a glacier may be divided, as developed by Benson (1961) and described by Paterson (1981). These zones vary in extent and importance from glacier to glacier, depending upon the regional and local climate.

Beginning from the upper most region of the glacier, there is the *dry snow* zone. Here there is no or negligible melting, even in summer. Continuing down slope this is followed by the *percolation zone*, where melting of snow does occur in summer. The line separating the two is called the *dry-snow line*. The melt water in the percolation zone, as the name suggests, moves into the snow and ice to some distance before it freezes. However, the snow does not become wet throughout its mass (Benson, 1961; Paterson, 1981).

This movement of melt water may complicate any ice core record by altering the concentrations of the various isotopes, dust particles and other components of snow which offer information about past climatic conditions. As this water refreezes, it releases latent heat which in turn warms the surrounding ice and snow. This is a considerable effect, as 1 gram of water releases enough heat to raise the temperature of 160 grams of snow by 1°C (Paterson, 1981). As the water percolates downwards, it may meet obstacles and form an *ice lens* or *ice layer*. The channels through which the water has moved may also freeze and form *ice glands*.

As the temperature of the glacier increases with decreasing altitude, there is the *wet-snow zone*. This zone has all of the snow deposited on it raised to its melting point. Here, some of the melt water will penetrate into the ice layers of previous years, which makes determining the mass balance for that season more difficult as those lower layers must somehow be included. This is separated from the previous zone by the *wet-snow line*.

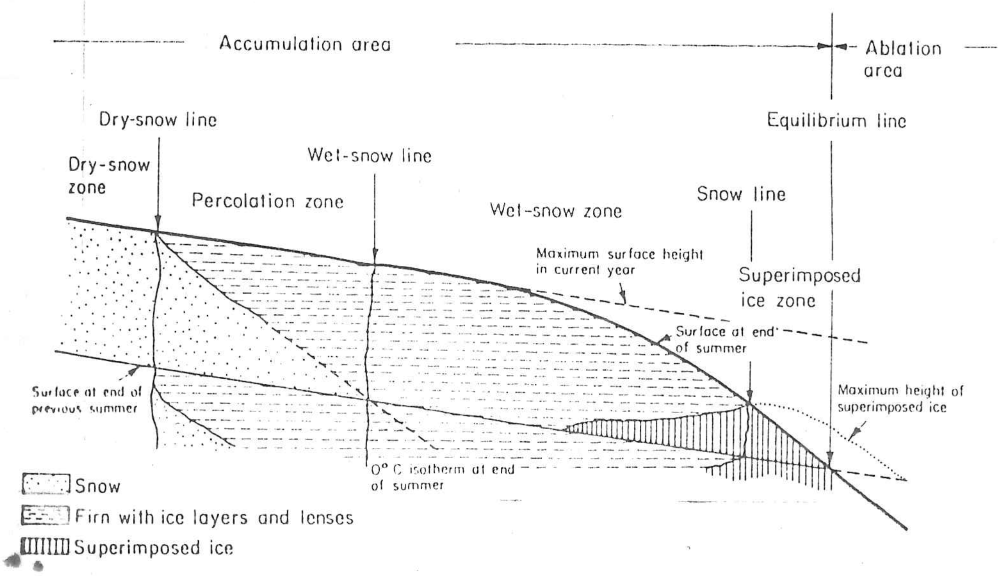


Figure 2.1 The zones of a glaciers accumulation area (Paterson, 1981).

The next area down slope is the *superimposed ice zone,* which is a continuous mass of ice formed by the refreezing of meltwater. Paterson (1981) restricts this term to that region where there is a net annual increment of ice. Sverdrup (1935) describes the formation of this zone. At the end of summer, the firn, which has a large amount of water contained in it, freezes. Next winter this layer will be covered by snow and this snow will in turn form the basis of another ice layer.

The boundary between the wet-snow zone and the superimposed ice zone is the *snow line* (also known as the *firn line, firn edge* and *annual snow line*). The lower boundary of the superimposed ice zone is the *equilibrium line*, below which is the *ablation zone*. The equilibrium line altitude is an important feature in determining how a glacier is reacting to a change in its climatic regime. There may, however, be situations where the superimposed ice zone does not form, hence the equilibrium line and the snow line would be the same. However for the Spitsbergen glaciers studied during this project, this was not the case and the two should not be confused.

2.3 The mass balance of glaciers

The *mass balance* of an ice surface is the algebraic sum of the accumulation and ablation (Paterson, 1981). It is the change in the mass of a glacier or ice mass expressed as volume per unit area. The mass balance of a given ice surface is dependent upon the accumulation and ablation of that surface (Paterson, 1981), or as Oerlemans (1991) describes it, the energy balance of the surface and the precipitation.

As mentioned in Chapter One, there has been a world-wide decrease in the extent of glaciers over the past 100 to 150 years. The reason for this decrease is generally considered to be the gradual increase during the last century in the global temperature. Oerlemans (1988), however, notes that the temperature change that has occurred, around 0.5K, is insufficient to describe the extent to which some glaciers have decreased their mass. This is evident by changes in the equilibrium line altitude. The examples used by Oerlemans (1988) to illustrate this are the glaciers Nigardsbreen (which has experienced a change in its equilibrium line altitude of 120 metres), Rhonegletscher (90 metres), Glacier d'Argentière (55 metres) and Hintereisferner (63 metres).

Kuhn (1989) determined that a change of 40 metres in the equilibrium line altitude would result from a temperature increase 0.6K, insufficient to account for the variations given in the previous examples. Altering parameters such as the net radiation budget and accumulation does produce variations in the equilibrium line altitude which are of the order of magnitude measured. This work, however, does not account for a number of factors and makes several assumptions which may not be justifiable. One of these is that changes in the mass balance profile are independent of altitude. Although this may be be the case for alpine glaciers (Kuhn, 1984), this is not always so. It also does not account for any feed-back effects, which may be significant.

An important characteristic of the mass balance of a glacier is the mass balance gradient, defined as:

$$F = \delta b / \delta z$$

where F is the mass balance gradient and b is the specific mass balance for altitude z. When precipitation is snow in the higher areas of the glacier, and rain in the lower parts, the result will be an increase in F. This is because the different forms of precipitation with altitude will result in altitude dependent changes in the glacier surface albedo. F also increases with the length of the ablation season (Kuhn, 1984).

2.3.1 Accumulation

Accumulation is the sum of the various processes which add mass to a glacier (Paterson, 1981). The dominant factor is precipitation in the form of snow. Other contributors include blown snow, rime, hoar frost, refreezing of rain and avalanches. The relative importance of each of these depends upon the climatology of the area. For example, rime will play a significant role in cool maritime regions such as the Antarctic Peninsula, while avalanches are more important around mountainous areas (Paterson, 1981).

A very important factor in accumulation is the surface air temperature. This determines whether precipitation will be snow or rain (Oerlemans and van der Veen, 1984). The temperature itself is in turn regulated by the precipitation as a result of a snow-ice-albedo feedback effect. If the surface temperature allows precipitation as snow, this will result in a higher surface albedo. This will then reduce the amount of radiation absorbed, further cooling the surface and thus increasing the possible area of snowfall. The opposite will occur if the precipitation were rain, as this will decrease the surface albedo with the reverse effect to that just mentioned (Barry and Chorley, 1987). When a glacier experiences a change from a positive mass balance to negative mass balance year, the melting of higher regions may reveal older ice with a lower albedo, hence further encouraging melting (Kuhn, 1984).

Another feedback effect will be when the altitude of the accumulation area is raised by successive seasons of increased snow. This will lower the temperature of the glacier's surface with an accompanying increase in the proportion of precipitation as snow. The effectiveness of this is partly dependent upon the topography of the region (Letréguilly and Oerlemans, 1990). This process is only important over long periods of time (hundreds to thousands of years) and hence is not a matter of concern for the time scales dealt with in this project.

The temperature of a region will also affect the humidity, which in turn relates to the amount of precipitation. A cooler atmosphere will have a lower capacity for water vapour, hence lower precipitation, as for example in the centre of the Antarctic Ice Sheet. However, mountainous regions have higher precipitation because of the cooling of rising moist air masses, another point which emphasises the importance of topography to an area's climate (Oerlemans and van der Veen, 1984).

2.3.2 Mass loss

Ablation is made up of the processes which cause a glacier or ice sheet to lose material (Paterson, 1981). This often occurs as a result of the melting of surface ice and snow. However, sublimation is significant around areas such as mountainous terrains (Paterson, 1981; Oerlemans and van der Veen, 1984). Around 70% of melting is a result of solar radiation. This figure varies from region to region, for example the figure for melting due to solar radiation for Lewis Glacier on Mount Kenya is 90% (Drewry, 1986). The remaining melting is a result of other processes, such as turbulent heat transfer and net longwave radiation.

Basal melting is another mechanism of mass loss. This is a result of the heating of the glacier's base by mechanical friction and stress and from the geothermal heat flux (Drewry, 1986). It also plays an important role in the loss of mass from floating ice shelves (Jenkins and Doake, 1991). Another way mass may be loss from a glacier or ice sheet is by calving, which is especially the case for ice shelves and sheets. In Antarctica this is the major cause of mass loss, making up 99% of total loss, while for the Greenland Ice Sheet this figure is around 50% (Oerlemans and van der Veen, 1984). As the glaciers studied during this project terminated inland, this need not be accounted for. However, other glaciers in the same area have calving as an important mechanism of mass loss and so it is worth noting that any mass balance study will need to account for loss by this process (Liestøl, 1988 and 1990).

An important requirement for surface ablation is an increase in the temperature of the surface of the ice. This may happen in a number of ways. Incoming radiation is one of the more important, namely solar radiation or longwave radiation from atmospheric components (e.g. water vapour, clouds). The resulting melt water may either run off, which is considered the usual case, evaporate or percolate into the ice mass, refreezing and hence warming the upper layer of the ice mass further. This raises the topic of the heat conduction of a glacier's surface, where ice is more conductive than snow (Oerlemans and van der Veen, 1984).

Other meteorological processes which assist in raising the ice mass surface temperature involves the atmospheric boundary layer and the motion of warm, moist air masses from lower latitudes. For high latitude regions, the radiation balance at the surface of a glacier or ice sheet is generally negative, resulting in a stable boundary layer below an inversion, as shown in figure 2.2. This layer experiences little vertical motion and the transfer of sensible heat is negligible. During the day this layer heats up as the radiation balance of the glacier surface becomes more positive, with a resulting rise in temperature at the surface and the start of convective motion (Oerlemans and van der Veer, 1984). If this process is strong enough, the result is the melting of surface ice and snow.

The intrusion of warmer, moist air masses from the lower latitudes is more efficient than the previous process at raising the temperature of the air above a glacier's surface, especially when it is snow covered (figure 2.3). The high winds which accompany such intrusions reduce the inversion layer, allowing significant sensible heat transfer (Oerlemans and van der Veen, 1984). These air masses are generally rising (warmer, hence less dense than the cooler high-latitude air it is displacing), with extensive cloud formation. This will result in an increase in the longwave radiation emission to the glacier's or ice sheet's surface, with the associated rise in surface air temperature.

2.4 The energy balance of a glacier surface

The energy budget of a glacier surface is very important to its the mass balance. Energy is transferred to an ice mass by solar radiation, longwave radiation from the atmosphere and clouds, turbulent heat flux and the refreezing of already melted water. It is lost by the emission of longwave radiation and the evaporation of water (figure 2.4).

The solar irradiance is a function of the time of day and year, in effect the position of the sun relative to the surface of the glacier. It is in the form of direct and diffuse solar radiation which is a result of the solar radiation's interaction with the various atmospheric components. These include clouds, the main atmospheric gases, water vapour, ozone, aerosols and dust (Paterson, 1981). Clouds reflect a proportion of the solar radiation back into space, the amount depending upon the fraction of the earth covered by cloud and the cloud type. This refers essentially to the albedo and thickness of the clouds. The average albedo of clouds varies from 0.9 for cumulonimbus to 0.4 for cirrus (Barry and Chorley, 1987).

Another important factor is the surface albedo. Albedo is the fraction of radiation reflected off a surface relative to that incident upon it. Typical values for glacier surfaces are presented in table 2.1. Factors which affect the albedo of an ice or snow surface are the size of the crystals, zenith angle

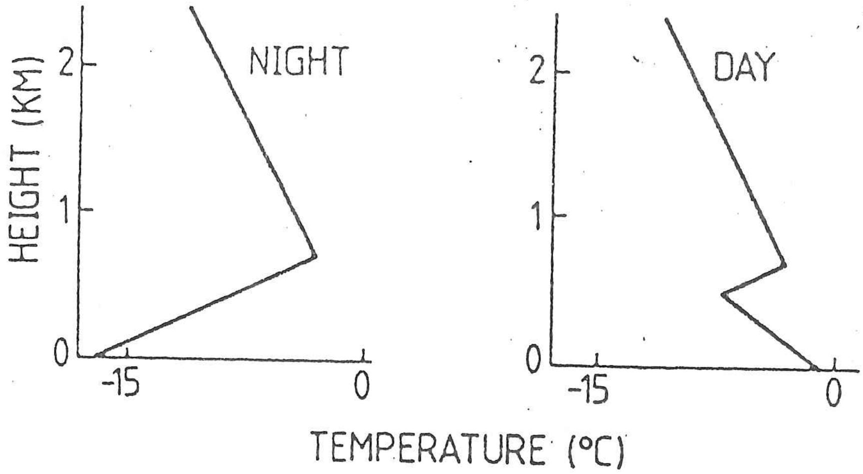


Figure 2.2 The stable boundary layer which results over an ice sheet. During the day, the increasing positive nature of the energy balance at the surface results in an increase in the temperature of the boundary layer (Oerlemans and van der Veen, 1984).

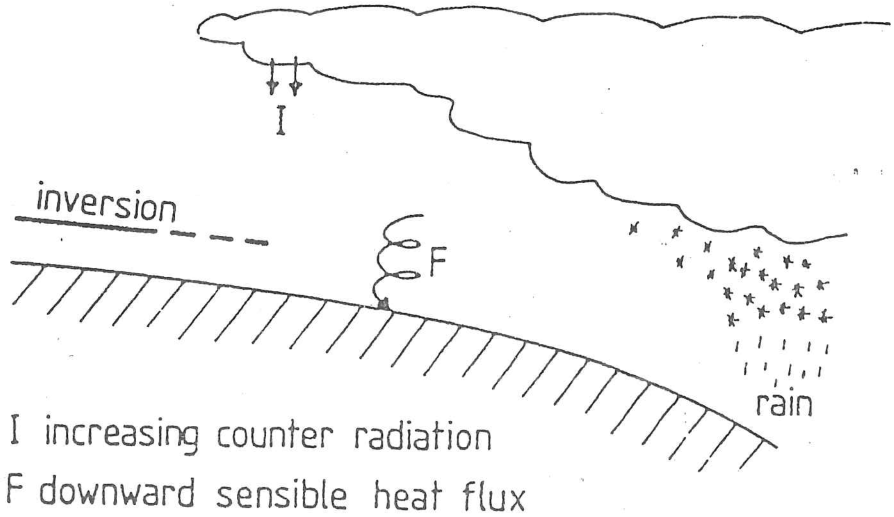


Figure 2.3 The intrusion of warm, moist air from lower latitudes which increases the temperature of the boundary layer near the ice mass surface (Oerlemans and van der Veen, 1984).

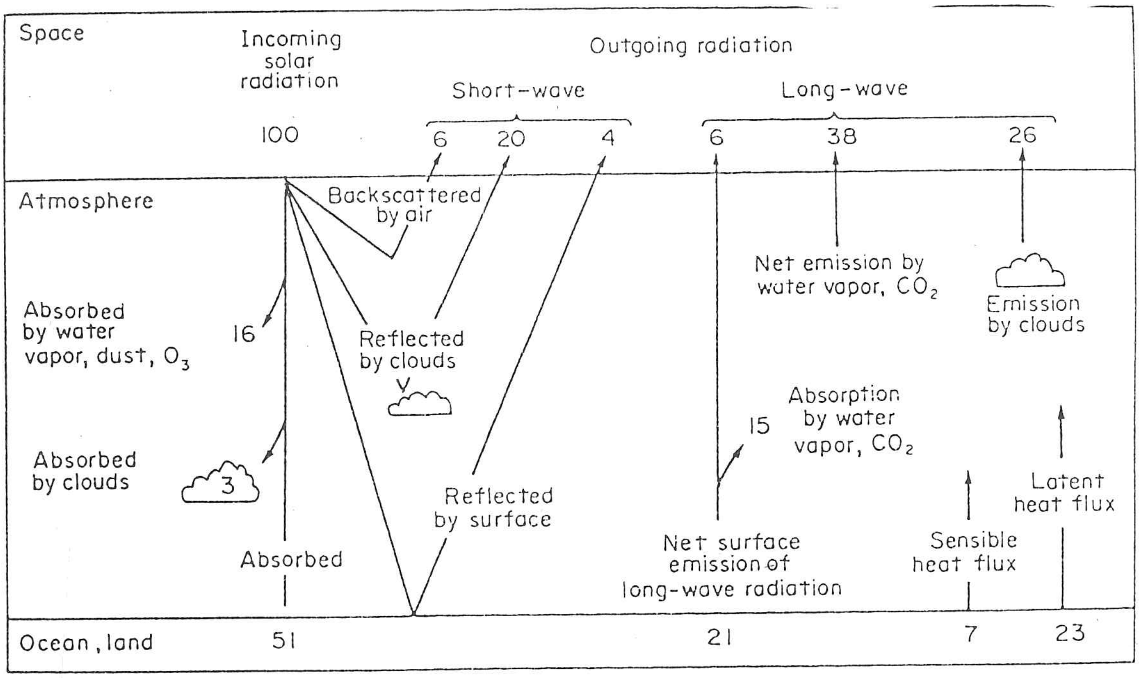
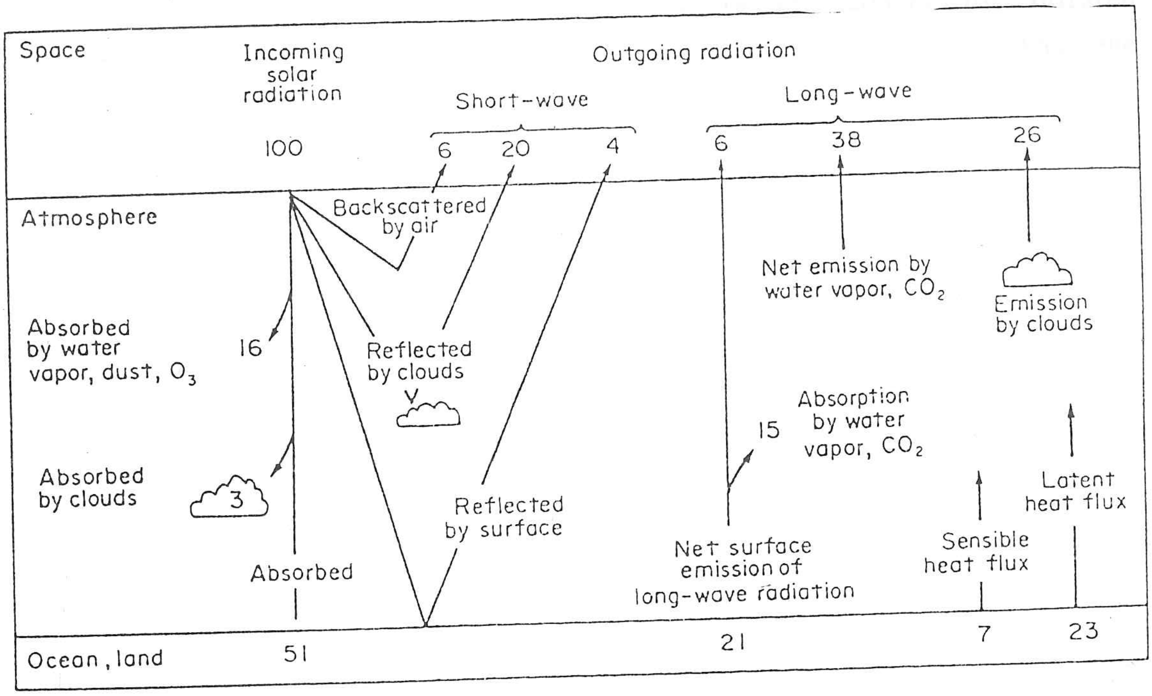


Figure 2.4 The radiation balance of the surface of the earth (Paterson, 1981).

New snow		0.68 - 0.95
old snow		
	clean	0.55 - 0.80
	dirty	0.45 - 0.60
firn		
	clean	0.50 - 0.65
	dirty	0.25 - 0.50
ice		
	clean	0.34 - 0.42
	dirty	0.18 - 0.34
	very dirty	0.11 - 0.18

Table 2.1 Albedo values for snow, firn and ice (Oerlemans, in press)



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of the sun, snow thickness and age and the presence of impurities. Each of these factors has a varying effect over the different parts of the spectrum. For example, the size of the ice grains alters the albedo for infrared radiation, while the visible is affected by snow depth. As ice gets older, the crystals enlarge, which explains the decrease in albedo as described above (Warren, 1982).

Impurities in snow in the form of soot resulting from natural or anthropogenic events, and volcanic ash, may have an important role in decreasing the surface albedo. Clouds affect the albedo of snow in that they alter the solar radiation, producing more diffuse light which snow and ice has different responses to. They also affect the spectral distribution of the incident radiation, with usually an increase in the effective albedo (Warren, 1982).

The longwave component of the energy budget is in two parts, that from the clouds and from the clear sky. The clear sky fraction results from the absorption and re-emission of solar radiation by the atmosphere. Clouds absorb solar radiation and re-emit it as longwave radiation, thus helping to counteract the reduction in solar radiation. They act as black bodies (Paterson, 1981) and when formulating the effect of the clouds in the energy equation for the model, the parameters which are of interest are the temperature and height of the cloud base. There will be occasions when the cloud base is warmer than the surface of the glacier, with the resulting net flux of long wave radiation from the cloud to the glacier surface (Paterson, 1981). Clouds also help to retain longwave radiation re-emitted from the earth. The outgoing longwave is limited to 316 W m^{-2}, which is that resulting from a surface at the melting point of ice. The surface of a glacier is taken to act as a black body (Paterson, 1981).

Two other processes which transfer energy to and from a glacier surface are the sensible and latent turbulent heat fluxes. The sensible flux is proportional to the difference in temperature between the air and glacier surface and transports heat by conduction and air motion. The latent heat flux results from the evaporation and condensation of water vapour onto the glacier surface (Paterson, 1981). The amount of heat transferred in these processes increases with turbulence (Paterson, 1981). Glacier winds have mixed effects on these fluxes, dependent upon the size and climate of the glacier (Ohata, 1989). In dry regions, the winds have little effect on the net turbulent heat, as a loss of the latent heat is corrected by a gain in the sensible heat. In humid semi-dry climates, the turbulent component is low if the temperature is low but increases with temperature. The strongest effect of glacier winds is on the ablation area of maritime glaciers.

It is useful to consider the difference between how ice or snow surfaces will behave compared with a surface of soil or rock when climatic conditions result in a change in the energy budget of that surface. If there is an increase in the incoming radiation on a non-frozen surface, the result will be an increase in the surface temperature. This will in turn result in an increase in the outgoing longwave radiation and turbulent heat flux until a new equilibrium is reached (Oerlemans, 1988). This new equilibrium may not be very different from the initial conditions. However, the temperature of ice or snow surfaces cannot increase above the melting point and the extra radiation is thus utilised in the melting of the ice and snow. Similarly, for the turbulent energy flux, when the air increases in temperature with the glacier surface's temperature remaining the same, the glacier will in effect absorb heat from the atmosphere (Oerlemans, 1988).

Chapter Three
Determining the glacier mass balance

This chapter will deal with the ways in which the mass balance of a glacier is determined, including computer modelling methods. First, there is a discussion covering the field methods used, taking for examples the measurements carried out on the Spitsbergen glaciers studied during this project. This will be followed by a description of how mass balance may be found from the correlation of measured balances and other more readily available data such as: 1) meteorological parameters from nearby weather stations and 2) the equilibrium line altitude. After this the mass balance models are discussed, with particular attention being paid to the model used in this project.

3.1 Field studies of the mass balances of glaciers

The gathering of field data is the most time consuming aspect of glaciological research. It is also the lack of such data which limits the testing of computer model results with actual measurements. For example, Liestøl (1982) comments that during 1981, bad weather made it impossible to make adequate snow accumulation measurements on the glacier Storbreen in Norway. It is also noted in Liestøl (1986) that during the 1983-84 season, there was a mild period in mid-winter around the Kongsfjorden area in north-west Spitsbergen, where the glaciers studied during this project are located. Some rainfall was recorded which resulted in a hard crust of ice forming on the glacier surface, making snow depth measurements more laborious.

The usual way to determine the net balance of a glacier is to make measurements at a series of points distributed on the glacier's surface (Paterson, 1981). These points need to be representative of the glacier as a whole so that an accurate idea of the balance conditions is gained. What must be measured is the mass of ice and snow which has accumulated during a balance year and remains at the end of the year, and the mass lost from the ablation area (Paterson, 1981). The net balance of the glacier is found from these two quantities averaged over the year. Winter balances may also be measured, with the summer balances determined from this and the net balance. These values are useful in determining any correlation between the glacier's behaviour and meteorological data at nearby weather stations.

Measurements of the mass balances of glaciers are done by the use of stake readings, pits and ice cores, snow depth and density measurements. Pits and drill cores are used to determine the net balance in the accumulation area. This means the previous summer's surface must be identified. It is usually marked by a layer of dirt or dust in the firn, or a change in the ice density, hardness or grain size. Melt water may percolate some distance into the glacier, refreeze and form layers of ice which may mix with the previous year's deposits (Section 2.2). This is especially the case in the superimposed ice zone, an area where pits are of little use (Paterson, 1981). This process also makes the identification of the annual layers more difficult.

Stakes are used to determine the balance in the superimposed ice zone and in the ablation zone. They are placed in holes drilled into the ice, the resulting balance being the difference between the top of the stakes to the ice surface at the start and end of the season (Paterson, 1981). Ice core drilling is used to determine the thickness of superimposed ice which, for most Spitsbergen glaciers, is a significant proportion of the year's accumulation (Liestøl, 1986).

For the glaciers studied, Austre Brøggerbreen and Midre Lovénbreen, these activities are started in May for accumulation measurements, with ablation being measured at the stakes in the middle of September. Pits for snow density measurements are dug at different altitudes, with snow depths surveyed along lines crossing the glacier (Liestøl, 1982). Austre Brøggerbreen and Midre Lovénbreen usually have negative annual mass balances. This means that the snow from the previous season would be nearly all gone, except at the highest regions. Snow depth could therefore be measured to an ice layer over almost the entire glacier, which would identify the winter accumulation with certainty (Liestøl, 1983). Because balances are not continuously recorded throughout the year, simply at the end of the winter and summer seasons, the values obtained for the yearly accumulation and ablation are not necessarily correct. However the resulting net yearly mass balance would not be affected.

There are other techniques that may be used to determine glacier mass balance. For example, hydrological methods which determine the output of melt water from a glacier may be used. However, only the net balance of the whole glacier may be found this way (Paterson, 1981). This method requires that the precipitation be known, as well as any evaporation or sublimation. Aerial and satellite imagery may also be used to determine the extent of a glacier from year to year, any changes noted and the volume determined.

3.2 Correlation of mass balances with other parameters

As mentioned, field surveys of glaciers are time consuming operations. An alternative approach is a model which determines the mass balance of a glacier using already recorded data and the correlation between this information and other parameters, or even from data gained from scaled down field operations. Reynaud *et al.* (1986) note this by offering two methods of mass balance measurement, both of which utilise the assumption of Lliboutry (1974) that interannual variations in a glacier's mass balance are independent of altitude.

The first is a *simplified linear model,* which gives the fluctuations in a glacier mass balance utilising a limited array of stakes arranged in profiles across the glacier. The balance is found at each stake position, with the mean found for each profile. The average balance variation is then found from these mean profile values. It does not give the total mass balance of a glacier, but is useful in that it monitors any fluctuations in the mass balance and may be used as a scaled down operation after a more complete series of surveys. The other method uses the *continuity equation* and requires the topographical survey of two glacier cross sections. This allows the mean mass balance of a section of a glacier to be found by the use of the continuity equation which relates the flux of ice through the cross sections.

The equilibrium line altitude is one of the more important parameters when investigating the responses of a glacier to climatic change. Braithwaite (1984) describes how it may be used to determine the mass balance of a glacier. This would be useful in that determining the equilibrium line altitude requires less field work, as only stakes in the appropriate part of the glacier would be required, as opposed to its entire area. Aerial or satellite imagery may also be used.

What is required is a time-series of mass balance and equilibrium line measurements. The expression relating mass balance and equilibrium line altitude would read as (Braithwaite, 1984):

$$b_t = \alpha (E_0 - E_t) \qquad 3.1$$

where b_t is the balance for a given year t, α is the effective budget gradient and E_0 and E_t are the equilibrium line altitudes when the balance is zero and at year t. A difficulty arises because an assumption made is that α is constant over time. This may not be the case because there is variation from year to year and with location about the glacier surface. Satisfactory results were gained, however, using this method for glaciers which had some available data. Efforts were made to estimate the mass balance of glaciers with no previous information, making use of certain assumptions concerning the likely position of the equilibrium altitude and the balance gradient. These proved to be unsatisfactory because of the large errors involved (Braithwaite, 1984).

Hydrometeorological models may also be used. These are simple statistical models that relate accumulation with winter precipitation, ablation to summer air temperature and other combinations of climatic parameters. They are useful in that they correlate climatic conditions at weather stations with the mass balance conditions of the glacier in question. Two examples are given by Hagen and Liestøl (1990) and Lefauconnier and Hagen (1990). These papers dealt with the statistical analysis of the mass balance of glaciers in Svalbard, amongst which were the glaciers modelled in this project, Austre Brøggerbreen and Midre Lovénbreen. It was found when correlating the mass balance of Austre Brøggerbreen with the climatic data at Ny-Ålesund (Section 4.3) that the best results came from comparisons between mass balance and positive summer and autumn temperatures, combined with winter precipitation (Lefauconnier and Hagen, 1990).

3.3 Mass balance models

Numerical models of glaciers depend upon the surface mass balance regardless of their degree of sophistication. Modelled studies of the climatic effects on glaciers may be carried out by using two assumptions: 1) that the mass balance changes are independent of altitude, as proposed by Lliboutry (1974), or 2) that the equilibrium line moves up and down, leaving the actual profile unchanged (Oerlemans and Hoogendoorn, 1989; Oerlemans, 1991).

The situation for the first assumption is valid for some glaciers, notably alpine glaciers (Kuhn, 1984). The balance of a point on a glacier as equated by Lliboutry (1974) is given by:

$$b_{jt} = \alpha_j + \beta_t + \varepsilon_{jt} \qquad 3.2$$

where b_{jt} is the balance at a point j and time t, α is the balance dependent on the position on the glacier, β is the balance which is dependent on time but independent of position, hence altitude, and ε is the error associated with time and position (Reynaud *et al.*, 1986).

An alternative is to adopt a more process-oriented approach, as has been done in this project. This involves consideration of the surface energy balance, either by a technique such as the degree day method or by defining the components of the surface energy budget themselves (Oerlemans, 1991). For example, the ablation method of Ambach (1989) was concerned with determining the interannual variation in the equilibrium line altitude. As ablation and accumulation at this position are equal, the resulting heat balance equation would be:

$$\tau_0 H_0 = k L C_0 \qquad 3.3$$

where τ_0 is the number of ablation days, H_0 is the average energy flux during an ablation day, k is a constant accounting for the combination of snow and superimposed ice, L is the latent heat of melting of ice and C_0 is the annual accumulation at this point designated by the subscript o which refers to it being the equilibrium line. The left hand side of this equation denotes the energy flux into this region while the right hand side is the amount of energy required to melt all of the snow and ice that has accumulated. The perturbations in the equilibrium line altitude become apparent when the equation is expanded to taken into account the altitudinal variations in the various parameters (Ambach, 1989).

Oerlemans (1988) suggests that at least two meteorological quantities are required to act as forcing for a glacier model. These are the air temperature and the radiation budget of the glacier's surface, essentially the quantities which are worked with in the model used in this project.

3.4 The model used in this project

The model used in this project is that of Oerlemans (1991). This model calculates the energy balance of a ice/snow surface. This section will draw mainly from Oerlemans (1991) and Oerlemans (in press). As the version of the computer code provided by Oerlemans (pers. comm.) had some minor modifications to that used in Oerlemans (1991), especially to some of the constants, the formulation of the program during this project used will be presented as opposed to that expressed in the above publications.

3.4.1 The model energy balance

The energy balance of a glacier surface may be written as:

$$B = Q(1 - \alpha) + I_{in} + I_{out} + F_s + F_l \qquad 3.4$$

where Q is the solar or shortwave radiation, α is the surface albedo, I_{in} and I_{out} are the incoming and outgoing longwave radiation, and F_s and F_l are the sensible and latent turbulent heat fluxes (Oerlemans, in press). The budget for the surface using this model divides the energy into four components; solar radiation, terrestrial radiation, turbulent energy fluxes and the refreezing of melt water. What is not accounted for in this model is the geometry of the glacier surface.

Solar radiation The solar radiation reaching the top of the atmosphere is found from the method of Walraven (1978). This radiation is attenuated by absorption and scattering, taking into account cloudiness, solar zenith angle and surface elevation.

The shortwave radiation is expressed as:

$$Q = S\,T_a\,T_c \sin\gamma \qquad \text{(when } \gamma > 0) \qquad 3.5$$
$$Q = 0 \qquad \text{(when } \gamma < 0)$$

$$S = 1353\ \{1 + 0.034 \cos(2\pi n/365)\}\ \ W\ m^{-2} \qquad 3.6$$

where S is the solar constant, n is the day number (were January the first is 1), γ is the solar elevation angle, T_a is the transmissivity of the atmosphere due to scattering and absorption by air molecules and aerosols, while T_c is the transmissivity constant for clouds. The transmissivity values are found by:

$$T_a = (0.79 + 0.00024\ h)\ \{1 - 0.009\ (90 - \gamma)\} \qquad 3.7$$
$$T_c = 1 - (0.41 - 0.000065\ h)\ \eta - 0.37\ n^2 \qquad 3.8$$

where h is the altitude above sea level in metres, and η is the fraction of the sky covered with cloud.

In the model, the terms S, γ and the last term in equation 3.7 are all included in a file which is read by the program. This file contains the appropriate values for each day for the latitude of the glaciers studied (Oerlemans, pers. comm.). The difference between diffuse and direct solar radiation is not accounted for. The solar zenith angle is also not accounted for when calculating T_c. This may result in the solar radiation at the surface being underestimated when there is a high surface albedo and scattered cloud (Oerlemans, in press).

Surface albedo This value is internally generated and is dependent upon the presence of snow, distance to the equilibrium line and the accumulated melt during the ablation season. This is given by the equations:

$$\alpha = \max\ [0.12;\ \alpha_{sn} - (\alpha_{sn} - \alpha_b)\ \exp(-d/200)] \qquad 3.9$$

$$\alpha_b = 0.115\ \mathrm{arctg}(h - E + 300)/200 + 0.43 \qquad 3.10$$
$$\alpha_{sn} = 0.75$$

where, α, α_b and α_{sn} are the resulting, background and snow albedos, d is the depth of snow (in mm of water equivalent), E is the equilibrium line altitude and h again is the altitude. The albedo is linked to the equilibrium line altitude, and because of this the program needs to be run for several cycles, each time redesigning the equilibrium altitude, until a stable solution is found (usually taking from three to five runs, Oerlemans (1991)).

Longwave radiation This component of the energy equation is divided between the outgoing radiation from the glacier surface and that incoming from the atmosphere. The atmospheric radiation is in turn in two parts; 1) the contribution from the clear sky and 2) from the clouds.

The outgoing radiation is limited to a value of 316 W m^{-2} (Section 2.4) since the surface must remain at or below the melting point. The surface of a glacier is taken to be a black body (Paterson, 1981). The incoming radiation is calculated using the formulation of Kimball *et al.* (1982). The radiation coming from the clouds is transmitted in the 8 - 14 μm spectral band, the so-called atmospheric window (Oerlemans, 1991 and in press).

The incoming radiation is expressed as:

$$I_{in} = \varepsilon_a \Theta_a^4 + I_{cl} \qquad 3.11$$

ε_a is the emissivity of the clear sky, Θ_a is the air temperature at screen level in Kelvin and I_{cl} is the radiation coming from the cloud base. The emissivity is expressed as:

$$\varepsilon_a = 0.7 + 5.95\text{x}10^{-7}\, e_a \exp(1500/(\Theta_a) - 2.5\text{x}10^{-5}\, h \qquad 3.12$$

where e_a is the vapour pressure (Pascals) and h altitude (equation 3.8). This formula is a compromise between that presented in Oerlemans (1991) and Kimball *et al.* (1982) as it accounts for the altitudinal variation.

The cloud contribution is expressed as:

$$I_{cl} = \varepsilon_{cl}\, n\, f\, \Theta_c^4 \qquad 3.13$$

where ε_{cl} is the emissivity of the cloud and f is a fraction of the blackbody radiation in the atmospheric band. Θ_c is the temperature of the cloud base in Kelvin and is found by applying a constant lapse rate, set for this area at -7^{o}C

km^{-1}, while n is the cloud cover (equation 3.8). ε_{cl} and f have been combined to a constant value of .25.

Turbulent fluxes The turbulent fluxes used by Oerlemans (1991) and Oerlemans (in press) are proportional to the difference between air and surface temperature and humidity. They were defined as:

$$F_s = C\,\Theta_a \qquad 3.14$$

$$F_l = 0.622\, L/c_p\, C\,(e_a - e_{as})/p \qquad 3.15$$

$$p = (1- .0001\ h)\ 1\text{x}10^5$$

where e_a is the vapour pressure found from the saturation vapour pressure and the relative humidity (equation 3.12), e_{as} is the saturated vapour pressure (Pascals), p is the atmospheric pressure (Pascals), L is the latent heat of melting, c_p is the specific heat and Θ_a is the temperature at screen height (in degrees Celsius) and C is the turbulent exchange coefficient. This is set for the version acquired from Oerlemans (pers. comm.) at 7 W m^{-2} K^{-1}. The saturated vapour pressure is found by:

$$e_{as} = 610.8\ exp(19.85\ (1 - 273.15/\Theta_a) \qquad 3.16$$

The exchange coefficient is assumed to be constant throughout the glacier extent. However Oerlemans (in press) notes that it tends to increase with decreasing altitude on large glaciers due to increased surface roughness and larger average wind speeds. The nature of the glacier's surface with regard to this parameter has not been accounted for. In the code obtained from Oerlemans (pers. comm.) the latent heat flux was not included in the heat budget equation. There were a number of exercises conducted in Chapters Five and Six (Sections 5.3 to 5.5, 6.2 to 6.4) which examined what influence this factor had on the resultant mass balances.

Refreezing of melt water The percolation of melt water plays a relatively important role in the energy of a glacier, especially those in Spitsbergen where superimposed ice contributes a significant proportion of the annual accumulation (Hagen, 1983). The amount of energy which goes into melting is defined by the energy balance and the following:

$$H_{ice} = B - R \qquad 3.17$$

$$R = B\, exp(\Theta_{ice})$$

where H_{ice} is the heatflux into the ice which warms the glacier's mass, R is the amount of energy available to melt the ice and snow, and Θ_{ice} is the temperature of the ice (Oerlemans, 1991 and in press).

3.4.2 The model mass balance

The mass balance of a ice mass is given by:

$$M = \int [\min (0, -B/L) + P]\, dt \qquad 3.18$$

where M is the mass balance, B is the energy balance at the surface, P is the rate of precipitation as solid (i.e. snow) and L is the latent heat of melting. The accumulation, melting and hence mass balance is calculated for each day.

There have been a number of simplifications in the energy calculations which have already been described. However there are other assumptions made, some of which will be tested in this project. The first concerns the precipitation as rain. It is assumed that precipitation will be snow when the temperature is below 2°C, above this temperature it is rain. When it is rain, it is assumed that all of it will run off. The reasoning behind this is that for the Greenland exercises conducted in Oerlemans (1991), the amount of precipitation which falls as rain is negligible.

The second assumption concerns how meteorological data is applied to the model. The daily temperature is determined by a sinusoidal equation which has the annual temperature mean and range defined. Precipitation is defined as being evenly distributed throughout the year such that it snows or rains every second day. Humidity and cloudiness are both assumed to be constant and are set to the annual averages. The turbulent flux co-efficient is constant throughout the year and over the entire glacier. The validity of some of these assumptions is tested in the studies presented in Chapters Five and Six (Sections 5.3 and 6.2).

A modification made to the model acquired from Oerlemans (1991) is that measured meteorological data may be used. The model also requires as input the values dependent upon the position of the sun, or the time of year (Walraven, 1978) for the calculation of the shortwave radiation component and the area distribution with altitude of the glaciers. The output is as balance profiles, total specific mass balance and equilibrium line altitude.

Chapter Four
The climate and glacial environment of Spitsbergen

This chapter describes the region and glaciers studied during this project. The general geography of Svalbard, especially Spitsbergen, will be discussed, covering the landscape, climate and glacial environment. This will be followed by a more detailed description of the Kongsfjorden area, including the meteorological station at Ny-Ålesund and the two glaciers studied, Austre Brøggerbreen and Midre Lovénbreen.

4.1 The Svalbard archipelago

Svalbard is the name given to the group of islands located between latitudes 74°N to 81°N and longitudes 10°E to 35°E (Hisdal, 1985). It is a region of great climatic variation caused by the confluence of very different air masses and oceanic currents.

This area (figure 1.1), is made up of four main islands; Spitsbergen, Nordaustlandet, Edgeøya and Barentsøya. There are also a number of smaller islands and groups of islands, for example Kvitøya, Kong Karls Land, Hopen and Bjørnøya. The total land area is approximately 63,000 square kilometres. Of this approximately 60% is covered by glaciers and ice caps. The amount each land mass is covered by ice varies from 99% for Kvitøya (Bamber and Dowdeswell, 1990) to none at all for Bjørnøya (Hisdal, 1985).

4.2 The island of Spitsbergen

4.2.1 The landscape

Spitsbergen is the largest island of the Svalbard archipelago. Much of it is very mountainous, especially in the north-east where the highest mountains, Newtontoppen and Perriertoppen (elevation 1717 metres), are located. The west coast is made up of steep, jagged mountains resulting from the metamorphic and eruptive geology of the area (Hisdal, 1985). The western and northern coastlines are extensively dissected by fjords, the two largest being Isfjorden to the west and Wijdeforden to the north (Hisdal, 1985).

The central region has large areas of plateaus formed from the flat strata of sedimentary rocks. The periods between glacials, and higher sea levels associated with them have left their mark in the moraines and coastal strandflats. Strandflats are low coastal plains partly covered by marine sediments which have been carved by the actions of sea and frost during interglacial periods. The best examples are along the western and northern coast of Spitsbergen (Hisdal, 1985).

Figure 4.1 shows the height in metres of the regional snow line for Svalbard. There are mountainous areas poor in glaciers such as Nordenskiöld Land between the fjords of Van Mijenfjorden and Isfjorden, and Andrée Land between Woodfjorden and Wijdefjorden in the north (Hisdal, 1985). These regions are considered to be the arid zones of Spitsbergen, as they are protected from moisture bearing winds by mountains and other land. As mentioned in Chapter Two, it is important that the snow line should not be confused with the equilibrium line altitude.

The actions of ice and frost are the dominant erosional forces in Spitsbergen, an obvious fact since just over half of Spitsbergen is covered by glaciers and ice caps. There is evidence in the form of various glacial erosional and depositional landforms which suggest that the ice extent was greater in the past (Troitskiy, 1981). These formations include glacial troughs and cirques, riegals, glacial moraines covering valley floors and marine terraces and polished rock surfaces with glacial striae (Troitskiy, 1981).

A common feature along the coastline of Spitsbergen (and all of Svalbard) are the presence of raised beaches. These resulted from the sea-level fluctuations and up-lifting of the archipelago which occurred during the early Holocene (Troitskiy, 1981, Troitskiy *et al.*, 1985). This uplift was caused by a combination of tectonic and glacial-isostatic processes. The height of these beaches in Spitsbergen varies from 3 to 40 metres, with some in other parts of Svalbard like Nordauslandet located between 40 and 70 metres. Dating of these beaches and the various moraines is carried out by

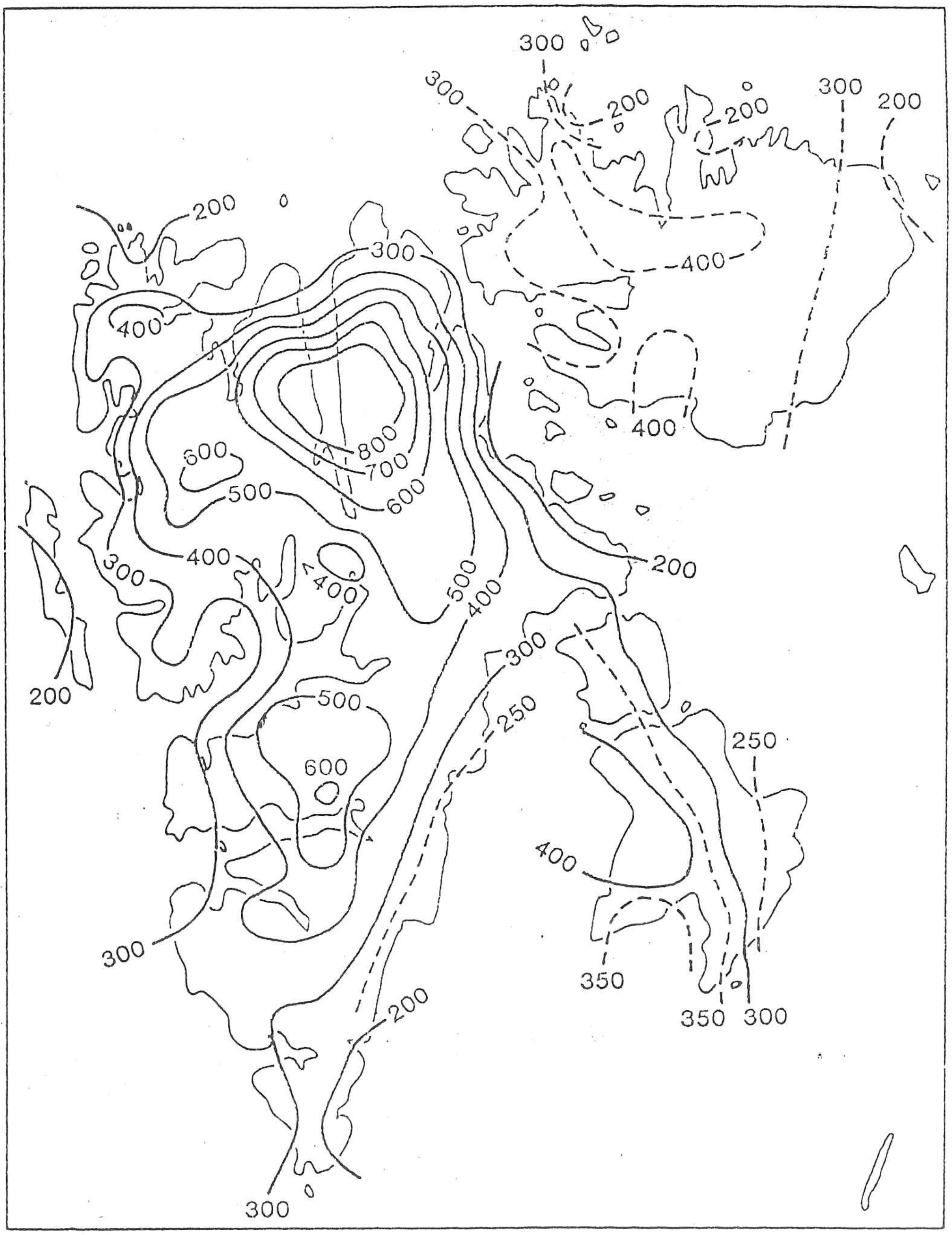


Figure 4.1 A map showing the height in metres of the climatic snow line (Hisdal, 1985).

several methods. These include the comparison of their stratigraphic locations, radioisotope dating of biological debris such as mollosc shells and driftwood, and lichenometry.

4.2.2 The climate

The climate of Svalbard is relatively mild considering its latitude, especially along the west coast. This mildness is largely due to the continuation of the warm Gulf Stream as the Norwegian Current through the Fram Strait. This region is the major pathway for warm southerly waters to enter the Arctic seas (Barry, 1986).

Using the Køppen system of climatic classification, Spitsbergen, at least at the meteorological stations, has a ET (Polar Tundra) climate. This implies that the mean air temperature of the warmest month is less than 10°C, with at least one month having a mean air temperature of greater than 0°C (Steffensen, 1982; Hanssen-Bauer *et al.*, 1990). For example, the average temperature for the coldest months at Isfjord Radio (78.1°N, 13.6°E) is -11.9°C. This compares with Isachsen, located at a similar latitude in the Canadian polar archipelago, which is around 20°C lower for the same period (Hisdal, 1985).

An important characteristic of the climate of Spitsbergen is its variability. This contributes to the difficulties encountered when studying the island's glacial history, especially when dealing with ice cores which often suffer from the problem of the refreezing of percolating melt water which may ruin the isotopic and other records (Kotlyakov *et al.*, 1990). These fluctuations in the climate are a result of two main factors.

The first concerns the circulation of the surrounding ocean currents. The Norwegian Current flows partly into the Barents Sea and partly towards the west coast of Spitsbergen, resulting in the northernmost region of open water in the Arctic. To the west, the cold East Greenland Current flowing southward along the coast of Greenland and a similar current along the east of Svalbard all contribute to the control of the climate of the region (Steffensen, 1982; Hanssen-Bauer *et al.*, 1990).

The second factor concerns the general atmospheric circulation. The large scale circulation of the North Atlantic is dominated by a low pressure regime over Iceland and the relatively high pressure over the Arctic Ocean and Greenland. This causes westerly to south-westerly winds between Iceland and Norway, the result being a movement of moist, mild air from lower latitudes northward. North of Svalbard, the circulation is anticyclonic with easterly to north-easterly winds. It is the temperature difference between these air masses which contributes to the variable climatic regime. The greatest variation occurs in winter when the temperature differences between the two air masses are most pronounced. Both of these factors are in turn coupled with the sea-ice extent, and between them contribute to the varied climate of this region.

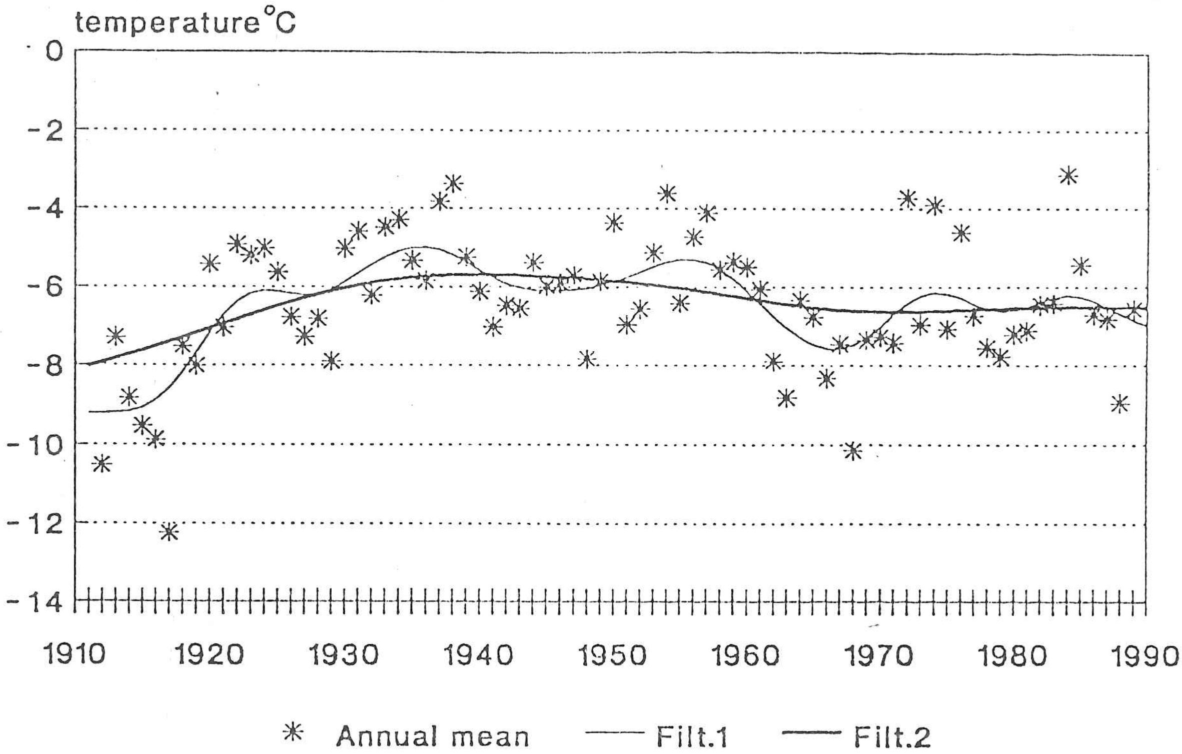
Precipitation in Svalbard is usually fairly low (Hagen and Liestøl, 1990) with values of between 200 to 400 mm yr^{-1}. The maximum occurs during August-September and minimum between April-June. It is difficult to obtain reliable measurements of precipitation because of the strong winds associated with it and snow drift. The actual amount and nature of precipitation also varies. For example snow is occasionally recorded in mid-summer, as is rain in winter.

The importance of identifying the climatic elements of the Svalbard region is apparent when attempting to determine if any changes are occurring to that region's climate. This is made more difficult by the lack of data. Table 4.1 shows the meteorological stations which have been, and are at present, in operation around Svalbard. It is apparent from this table that there is a problem with obtaining a continuous record with time of basic meteorological measurements. None of the present stations have a proper time series extending to earlier than the mid-1970's, and it was also at this time that the stations with the longest records, Isfjord Radio and Longyearbyen (figure 1.1), were shut down (Hanssen-Bauer *et al.*, 1990). However, by making use of data from the present station at Svalbard Lufthavn and from other stations (Hanssen-Bauer *et al.*, 1990) a continuous time series since 1912 has been constructed, as shown in figure 4.2.

This graph shows that, from 1912 to 1940, there was an increase in the annual mean temperature. Hesselberg and Birkeland (1940) mention this in connection with the decreasing extent of the glaciers, the increasing length

Isfjord Radio	Sept. 1934 - Sept. 1941 Aug. 1946 - June 1976	78° 04'N, 13° 38' E
Longyearbyen	1916 - 1923 (incomplete) Sept. 1930 - Aug. 1934 1935 - 1939 (incomplete) Oct. 1941 - June 1942 Sept. 1945 - Aug. 1946 Jan. 1957 - July 1977	78° 13' N, 15° 35' E
Svalbard Lufthavn	Aug. 1975 to present	78° 15' N, 15° 28' E
Ny-Ålesund	(see table 4.2)	
Svea Gruber	May 1978 to present (incomplete)	77° 54' N, 16° 48' E
Storøy	1908 - 09, 11 - 12. 14 - 15	76° 30' N, 16° 30' E
Akseløy	1898 - 99, 1900 - 01, 02 - 03, 04 - 05, 10 - 11	77° 42' N, 14° 50' E
Green Harbour	1911 - 1930	78° 02' N, 14° 14' E
Quade Hook	1912 - 1924 (incomplete)	78° 57' N, 11° 42' E
Andersonøy	1894 - 95	78° 20' N, 20° 44' E
Kap Lee	1904 - 05	78° 06' N, 20° 55' E
Hvalfiskpynten	1894 - 95, 1904 - 05, 06 - 07, 08 - 09	77° 30' N, 21° 00' E
Zieglerøya	1904 - 05	77° 20' N, 22° 02' E
Halvmåneøya	1906 - 07	77° 17' N, 23° 05' E
Nordostlandet	1944 - 45	80° 04' N, 22° 24' E
Hotellneset	1964 - ???	78° 15' N, 15° 33' E
Hopen	Nov. 1944 - July 1945, Nov. 1945 to present	
Bjørnøya	Jan. 1920 - June 1941, Nov 1945 to present	

Table 4.1 Meteorologial stations, past and present, of Svalbard (Steffensen, 1975; Hanssen-Bauer *et al.*, 1990).



Filter 1 a Gaussfilter with STD = 3 years

Filter 2 a Gaussfilter with STD = 9 years

Figure 4.2 Annual mean temperatures reconstructed for Svalbard Lufthavn 1912 - 1990 (Hanssen-Bauer *et al.*, 1990).

of the shipping season and some biological phenomena. For example, at this time bird species which normally migrated south for the winter did not do so. This was followed by a decrease in annual air temperature between 1950 to 1970. Since 1970 the annual mean temperature has remained roughly constant, although there has been a slight increase in spring temperatures while for the other seasons for the last 20 years they have been fairly constant (Hanssen-Bauer *et al.*, 1990).

With regards to what is expected in terms of climate change, present modelling studies indicate that any change to the global surface temperature brought on by an increase in greenhouse gasses will be amplified in the polar regions, especially the Arctic (Koster, 1991; Walsh, 1991). There are studies which show a hemispheric asymmetry in the changes between the Arctic and Antarctic, with the Arctic showing the effects of warming earlier than other regions of the Northern Hemisphere (Stouffer *et al.*, 1989).

This is supported by some observations of the sea-ice conditions in recent years. For example, Wadhams (1989) proposes a thinning of the Arctic sea-ice, while Gloersen and Campbell (1991) describe a decrease in its extent. However, Stouffer *et al.* (1989) comment that any temperature increase will be less around the North Atlantic because of a decrease in the thermohaline circulation, resulting from a lessening of the salinity of the upper oceanic layer because of increased runoff and precipitation. Walsh (1991) also notes that certain effects such as the ice albedo-temperature feedback are not fully understood, and this, coupled with the lack of reliable data, means that conclusive predictions cannot be made.

4.2.3 The glacial environment

The glacial environment of Spitsbergen is made up of ice caps, large ice masses divided into streams by nunatuks and mountain ranges, and smaller cirque glaciers, especially along the western coast. The glaciers of Spitsbergen, like many other regions of the world, are experiencing a decrease in their extent, evident from mapping and photographic records and mass balance studies (Liestøl, 1988, 1990). This shrinking is partially off set by the occurrence of frequent surging (Elverhøi *et al.*, 1980). Surge type glaciers appear to make up approximately 90% of glaciers in Svalbard with some advancing at any one time (Hagen and Liestøl, 1990). This fact is one of the more complicating aspects of the glaciology of Svalbard, especially when attempting to determine the relative ages of moraines.

The glaciers of Spitsbergen, and all of Svalbard, have experienced several fluctuations in their extent during the Holocene. This period has been divided by Troitskiy *et al.* (1985) into four stages; the Damesmorenen (7800 years B.P.), the Passdalen (4500 - 4000 years B.P.), the Grønfjorden (3000 to 2500 years B.P.) and the Treskelen or Little Ice Age (17th - 19th centuries). The minimum in the glaciation during the Holocene was around 5500 to 5000 years B.P.. During this time there was an increase in the precipitation which prevented the ice masses from disappearing completely (Troitskiy, 1981; Kotlyakov *et al.*, 1990).

The Holocene advances of the glaciers were in many areas most intensive during the Little Ice Age, during which their fronts overrode the deposits left by earlier advances (Grove, 1988). One of the peculiarities of the glaciation of Svalbard is that there are regions where this maximum occurred at the start of the Holocene. This is a similar situation to that which occurred in western North Africa and contrasts with that of Scandanavia, the Alps and Caucasus which experience a gradual decrease in the extent of their glaciers throughout the Holocene (Troitskiy *et al.*, 1985).

The glaciers of Spitsbergen have been studied by a number of methods, including gravity, radio echo soundings (ground and airborne) and seismic. Because many of the Spitsbergen glaciers are of the sub-polar type (Section 2.1) the use of radio echo sounding becomes difficult because of the absorption and scattering of the signals (Dowdeswell *et al.*, 1984).

Ice core studies are particularly useful when determining the past climatic conditions. The first ice core work on Svalbard was conducted during the Soviet glaciological surveys from 1965 to 1967 (Zagorodnov, 1988). This was followed in 1974 by more detailed studies covering the mass balance, hydrology, thermal history and regime, and the fluctuations and evolution of the glaciers. The large amount of melting which occurs on Spitsbergen glaciers may degrade some of the ice core record. However in the parts of Spitsbergen where the melt water does not penetrate very far, such as the Lomonosov Plateau (Gordiyenko *et al.*, 1981), the formation of ice bands helps to define individual years.

4.3 The Kongsfjorden area

The region of Spitsbergen where the glaciers studied, Austre Brøggerbreen and Midre Lovénbreen, are located is around the peninsula of Brøggerhalvøya, along the southern coast of Kongsfjorden (figure 4.3). The meteorological station of Ny-Ålesund, where the weather data used in this project were recorded, is also located here. For simplicity, the glaciers will be referred to as Brøggerbreen and Lovénbreen for the remainder of this thesis.

The glaciers which drain into Kongsfjorden cover an area of 1031 km^2. This is dominated by the glaciers at the head of Kongsfjorden which are divided into three calving faces; Kronebreen, Kongsvegen and Conwaybreen. To the north there is Bloomstrandbreen. None of the glaciers on the Brøggerhalvøya peninsula actually reach the sea (figure 4.3). The maximum altitude of the mountains along this peninsula is above 700 metres (Norsk Polarinstitutt, 1979). The maximum winter snow depths vary from approximately 1 metre at sea-level to 3 metres in the higher regions of the glaciers (Hagen, 1988).

4.3.1 The Ny-Ålesund weather station

Meteorological data used in this project are from the Norsk Polarinstitutt weather station at Ny-Ålesund. Details of this station are shown in table 4.2. Ny-Ålesund has the advantage of being close to the glaciers studied, approximately 3 to 5 kilometre distance. It is located on a flat tundra plain with the nearest mountains only around 1.5 kilometres away (figure 4.4).

There have in fact been two stations, one operating from 1969 to 1974 which was then moved 1.6 kilometres SSE. The first station was located on the axis of Brøggerbreen and as such was affected by the katabatic winds off the glacier (Lefauconnier and Hagen, 1990). The second station is located in an area whose winds are more representative of the area (Steffensen, 1982; Hanssen-Bauer *et al.*, 1990). The temperatures at the newer station are slightly lower during the spring and higher in precipitation. Figure 4.5 shows the average monthly temperature, precipitation, relative humidity and cloudiness from 1975 to 1989 for the station.

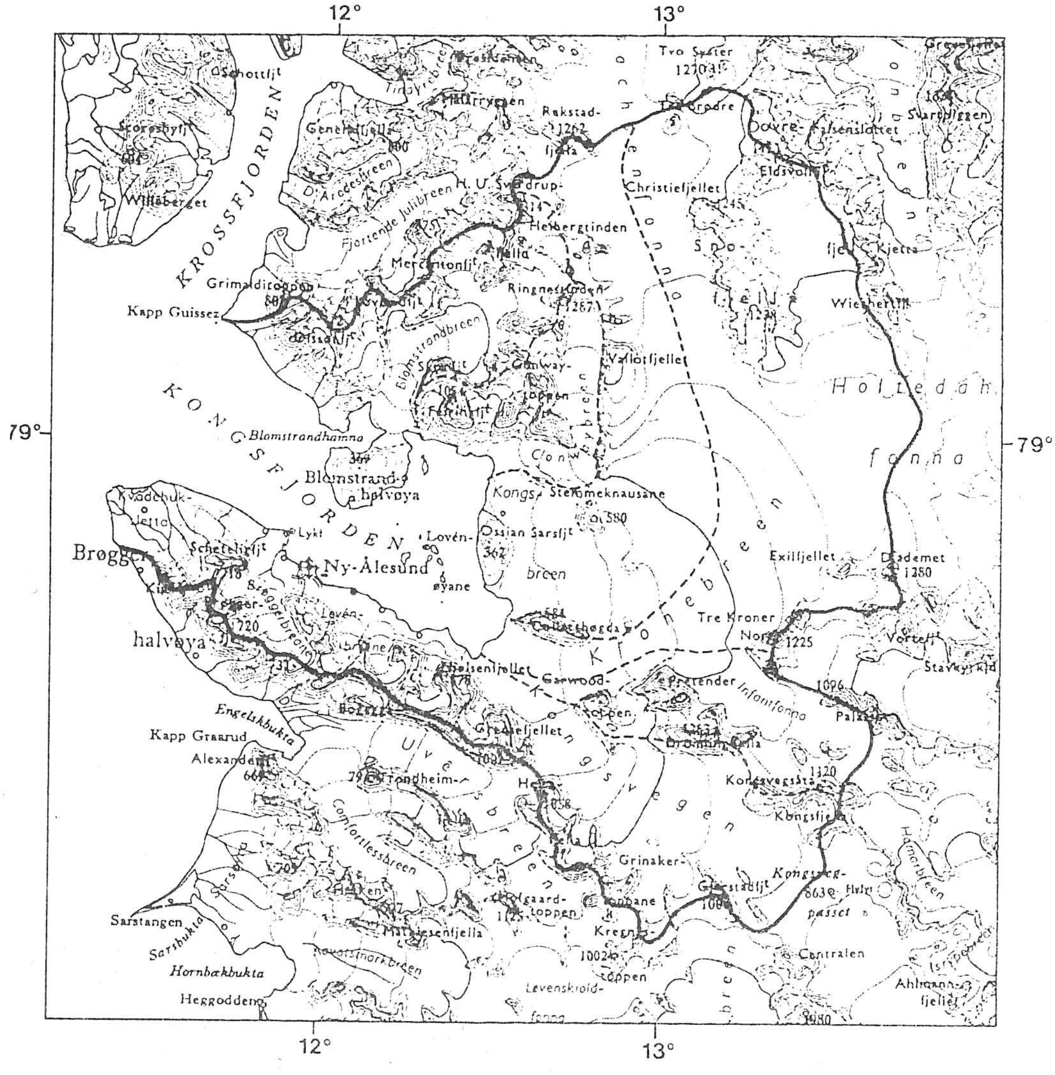


Figure 4.3 The Kongsfjorden area, north-west Spitsbergen (Liestøl, 1990).

Ny-Ålesund I

1961 - 1968 (incomplete)	
1969 - July 1974	
Position:	78° 56′N, 11° 53′E
Station height:	42 metres

Ny-Ålesund II

Aug. 1974 to present	
Position:	78° 55′N, 11° 56′E
Station height:	8 metres
Midnight sun:	18th April to 24th August
Polar night:	25th October to 17th February

Table 4.2 Some details concerning the meteorological station at Ny-Ålesund (Hanssen-Bauer *et al.*, 1990).

4.3.2 Brøggerbreen and Lovénbreen

Brøggerbreen and Lovénbreen (figure 4.4) are small cirque glaciers with areas of around 6 km^2 (Hagen, 1988). These glaciers are quite well studied, mass-balance observations of them being made by the Norsk Polarinstitutt since 1966 for Brøggerbreen and 1967 for Lovénbreen. Field studies of these glaciers include core drilling, stake measurements and radio echo soundings (Liestøl, 1988 and 1990).

During the period of mass balance measurements, only 1 year for Brøggerbreen and 2 years for Lovénbreen have shown positive mass balances, and only three seasons show values close to equilibrium (figure 4.6). This suggests that the glaciers are decreasing in mass, an observation backed by moraine deposits and photographs of the front of these glaciers from earlier this century (Hagen and Liestøl, 1987; Liestøl, 1988 and 1990; Lefauconnier and Hagen, 1990). This decrease in the extent in glaciers is connected with a world-wide phenomena as described in Chapter Two. This decrease is more obvious in the smaller glaciers of this area where they have lost sometimes over a half of their mass since late last century. Lefauconnier and Hagen (1990) note that there is a trend for the balances to become less negative as observed since the mass balance measurements of these glaciers started in the late 1960's.

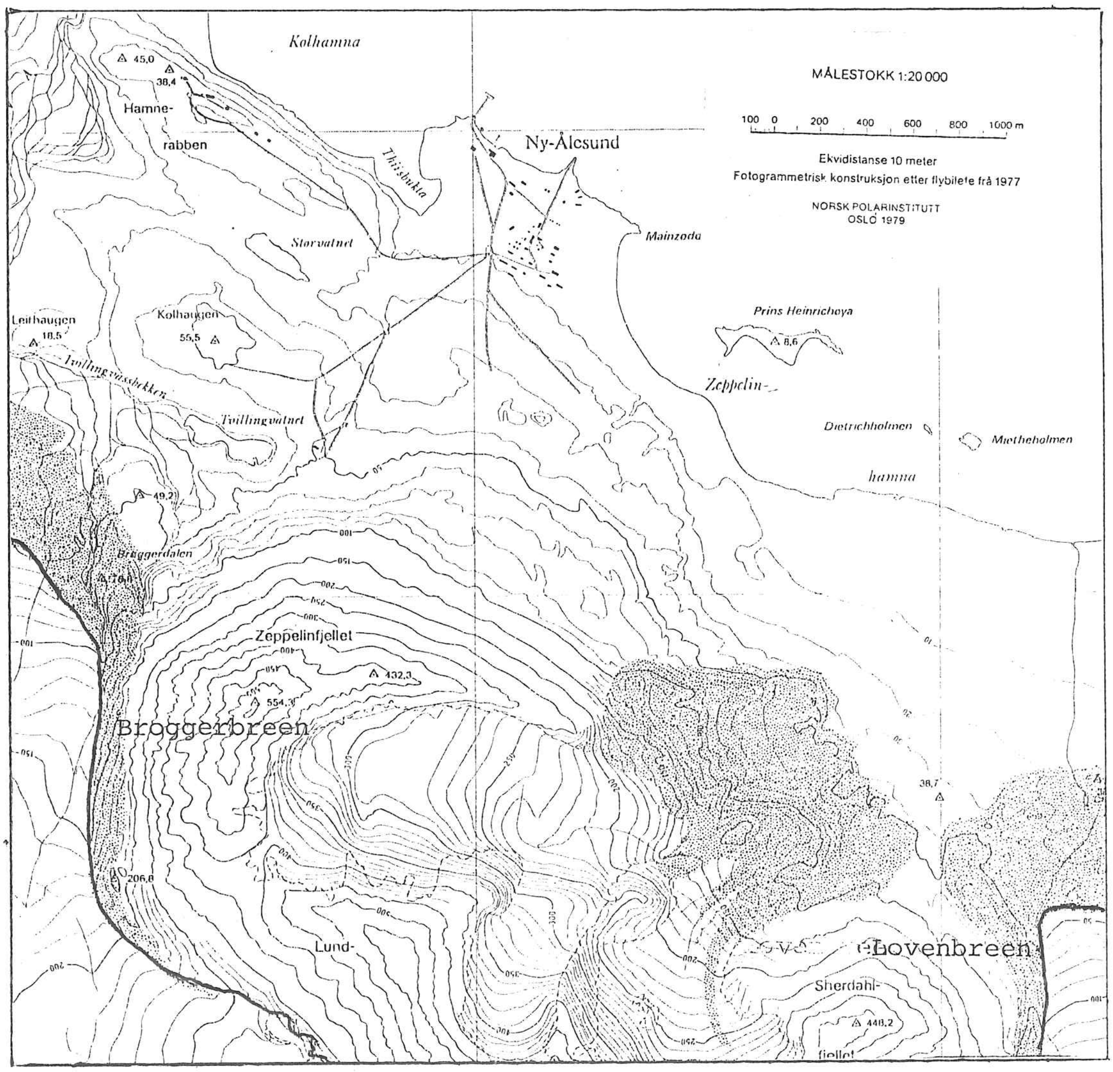
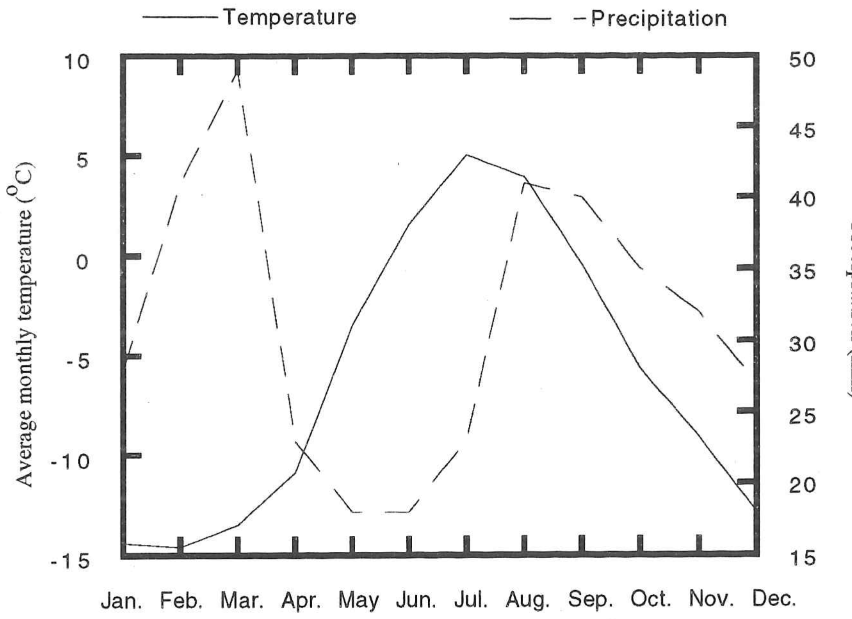
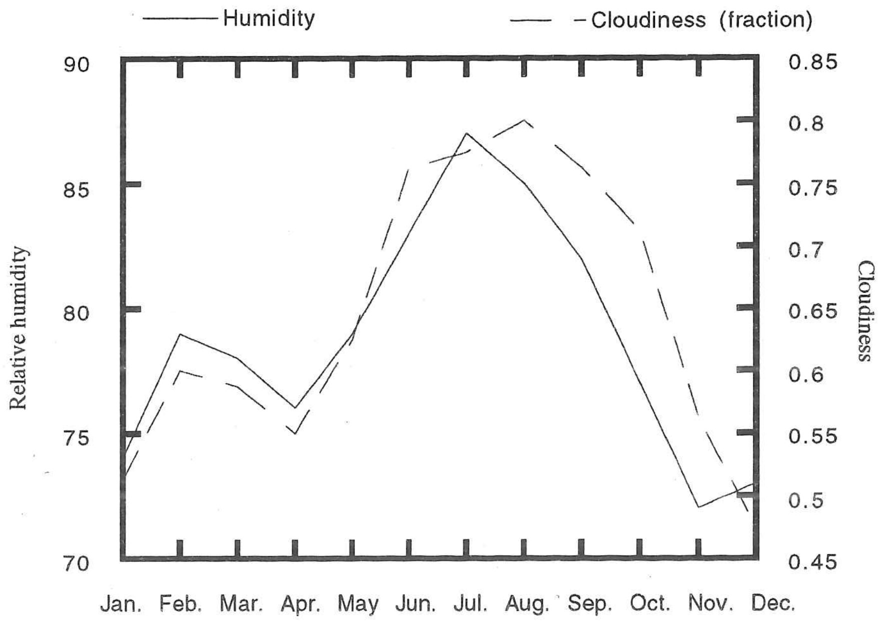


Figure 4.4 The glaciers Lovénbreen and Brøggerbreen, and the weather station at Ny-Ålesund (from Norsk Polarinstitutt, 1979).

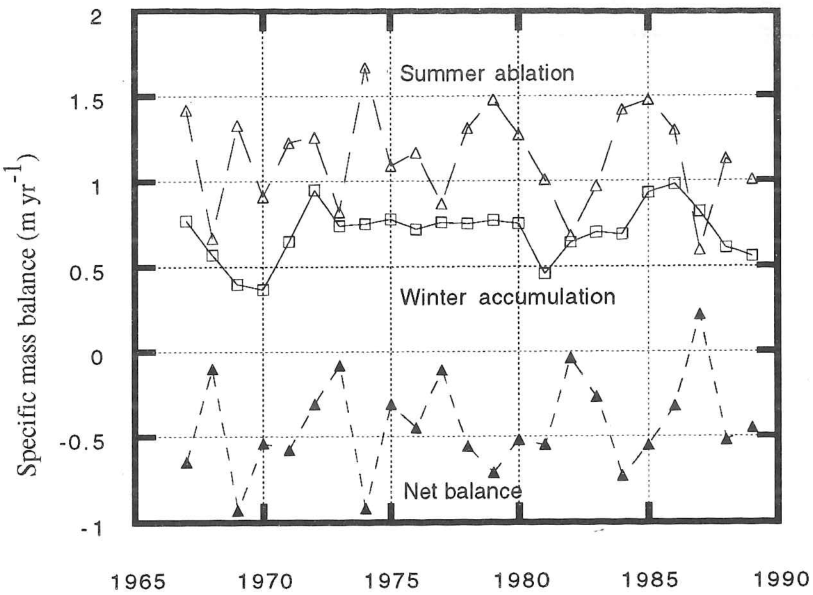


a) Average monthly air temperature and precipitation.

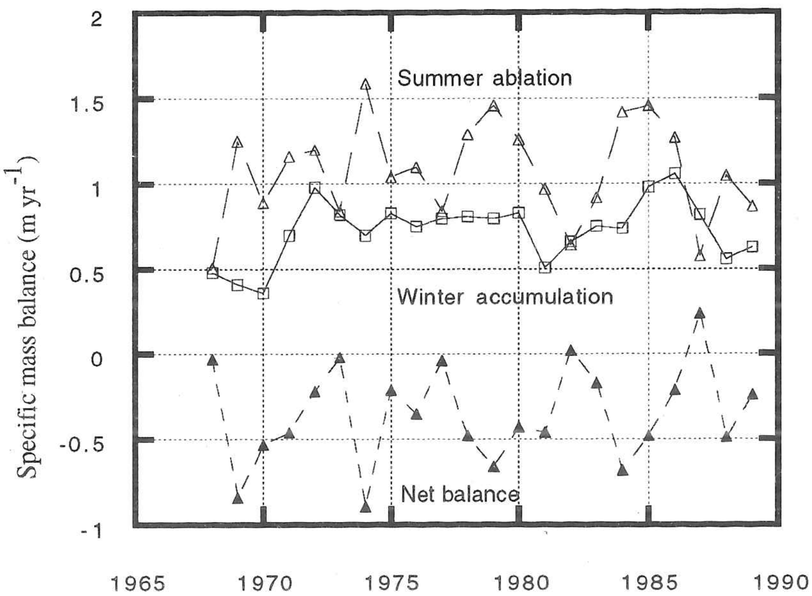


b) Average monthly relative humidity and cloudiness.

Figure 4.5 Average meteorological data for Ny-Ålesund between the years 1975 to 1989 (Hanssen-Bauer *et al.* 1990).



b) Mass balance of Brøggerbreen from 1967 to 1989.



b) Mass balance of Lovénbreen from 1968 to 1989.

Figure 4.6 Annual mass balances of a) Brøggerbreen and b) Lovénbreen (Liestøl, 1988 and 1990).

The balance values of the two glaciers are closely related because of their proximity. However, the mass balance of Brøggerbreen is usually less than that of Lovénbreen owing to the fact that a greater proportion of it is at a lower altitude (Hagen, 1988; Section 5.2). Shrinkage of the glaciers is more pronounced around the glacier tongues (Liestøl, 1980). There is good correlation between certain climatic parameters at Ny-Ålesund and the mass balances (Lefauconnier and Hagen, 1990). These parameters include the positive summer and autumn temperatures and winter precipitation (Section 3.2).

There is some year to year variation in the snow cover pattern of the two glaciers. However, this is almost always caused by variation in the rain/snow distribution with height in autumn and spring (Liestøl, 1980). Snow distribution is approximately doubled going from the lowest to highest points on the glaciers. A significant proportion of the accumulation of these glaciers is the formation of superimposed ice. This adds to the difficulty in modelling the mass balance of these glaciers.

A borehole at the front of Brøggerbreen has measured permefrost to a depth of 140 metres, making it likely that this frost goes beneath the glacier tongue and the sides. In the firn area, the temperatures down as far as the bedrock are at the pressure melting point, except for the upper few metres in winter. This is due to the fact that the snow is saturated with meltwater during summer, the heat capacity of water being high enough to prevent it from freezing in the winter, except for the upper most 5 to 10 metres near the surface (Liestøl, 1988 and 1990). This water supplies the groundwater which is below the permefrost. Water also finds its way along the bed of the glacier, forming icings at the front of glaciers in winter.

From radio echo soundings, it is found that Brøggerbreen has a depth of approximately 130 metres. It does not drain water while Lovénbreen, which is around twice as thick, forms large icings in winter. These icings are an indicator of the temperature regime of a glacier, whether it is totally below freezing point or of the sub-polar type with the accumulation area above the melting point, which both of the glaciers studied are.

Chapter Five
Mass balance modelling of Spitsbergen glaciers: Modelling procedures

This chapter will deal with the mass balance modelling of the Spitsbergen glaciers, Lovénbreen and Brøggerbreen. Preliminary tasks such as determining the area/altitude distribution of the glaciers will be outlined, as well as specifying which meteorological data from the weather station at Ny-Ålesund will be used. The first tests were sensitivity studies which show how perturbations in the various parameters varied the resulting specific mass balances. This was followed by each of the objectives specified in Chapter One. The first is a comparison between the Oerlemans (1991) method of applying meteorological data to the model and the use of measured climatic data. Next will be the mass balance modelling of Lovénbreen and Brøggerbreen for present day climatic conditions, and thirdly, modelling the mass balances of these glaciers for future warmer, and past cooler conditions. The results are presented in Chapter Six and Appendix Three.

5.1 Initial procedures

5.1.1 Determining the area distribution of the studied glaciers

The first requirement before the mass balance modelling of the Spitsbergen glaciers, Lovénbreen and Brøggerbreen, could be carried out was to determine their area distribution with altitude. This was achieved using a Mini-mop digitizer unit and a 1:20,000 scale map of the area covering the two glaciers (Norsk Polarinstitutt, 1979).

This map for most of the glaciers' extent was contoured at an interval of 10 metres. The glaciers were divided into altitudinal segments of 20 metres until a height of 500 metres, above which intervals of 50 metres were used. This was because the contour spacing above 500 metres was usually greater than 10 metres. Some sections of the glaciers below this height were also contoured at a coarser scale, in which case interpolation was carried out.

This was done by matching the interpolated lines with those bounding the segment of interest. Figure 5.1 shows the resulting area distribution of the two glaciers with altitude. This figure better describes the comment made in Section 4.3.2 about how the area distributions of the two glaciers differ. Lovénbreen, as can be seen, has a greater proportion of its area at a higher altitude than Brøggerbreen, hence the lower mass balances recorded for Brøggerbreen compared with Lovénbreen (Section 4.3.2, figure 4.5).

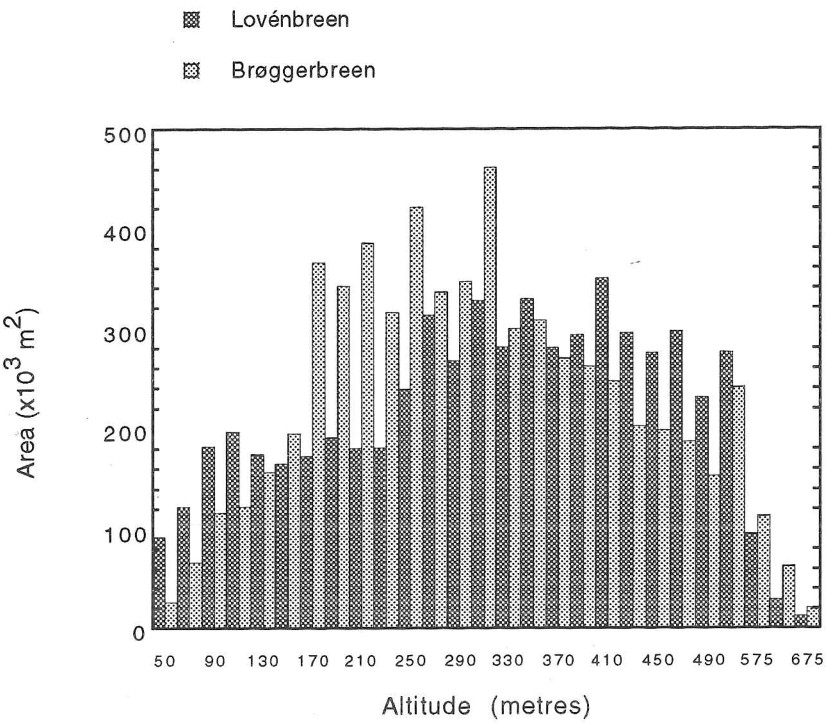


Figure 5.1 The area distribution with altitude of the glaciers Lovénbreen and Brøggerbreen (Norsk Polarinstitutt, 1979).

5.1.2 The meteorological data

The meteorological data used were obtained from the Norsk Polarinstitutt. This information consisted of the daily average temperature, precipitation, relative humidity and cloudiness recorded at the Ny-Ålesund weather station. The data were in the form of computer printouts and were entered into the computer manually. The data used was for the years 1983 to 1988 and it will be these seasons which will be modelled. The meteorological data used are presented in Appendix Two (figures A2.1 to A2.5).

The season of 1983/84 was the most negative of the years modelled (Liestøl, 1986). This was due largely to a mild period with some rainfall in mid winter. The 1984/85 and 1985/86 seasons were characterised by higher snowfalls (130 and 140% the normal for Brøggerbreen). However, this was countered by an increase in the ablation caused by warmer summers. This resulted in a more negative than usual mass balance in 1984/85 and a less negative balance in 1985/86 (Hagen and Liestøl, 1987). The 1986/87 season was a positive balance year for both glaciers (Hagen, 1988), the only one on record for Brøggerbreen. Lovénbreen has had one other positive balance year, 1981/82.

Corrections carried out on the meteorological data were for temperature change with altitude (Section 3.4.1), and for some of the modelling exercises, precipitation with altitude (Section 5.4). There were no corrections for cloudiness or humidity over the extent of the glacier. For cloudiness this may be significant, as Liestøl (1977) comments on the fact that cloud cover at the weather station, and between mid to high glacier, all differed.

5.2 Sensitivity studies

The first modelling experiments carried out were a series of sensitivity studies. The purpose of these tests was to examine how perturbations in certain parameters appeared to affect the model results. The parameters varied were the annual average temperature, annual precipitation, relative humidity, cloudiness, height of the cloud base and the glacier surface albedo. Some useful examples of other sensitivity tests are presented in Oerlemans and Hoogendoorn (1989) and Oerlemans (1991). Although not one of the specific aims of this project, it is important to know how much the results would be affected by changes in the various parameters.

The meteorological data were applied to the model in the same manner as Oerlemans (1991) (Section 2.4.2). The daily air temperature was calculated by defining the annual temperature range and the yearly temperature mean, and applying these to a suitable sinusoidal function. The precipitation was assumed to be constant and equally divided so that it either snowed or rained every second day. If the temperature was less than 2°C, it snowed, otherwise rainfall, which was assumed to run off. Relative humidity and cloudiness were constant throughout the year. The cloud base altitude was set to be 100 metres above the highest segment of the glacier and the albedo was as defined by Oerlemans (in press).

The climatic and glacier parameters were varied in the following manner; the annual average temperature by +/- 1 to 2^{o}C, the precipitation, cloudiness and relative humidity by +/- 10 to 20%, the cloud base altitude by +/- 100 to 200 metres, and the surface albedo of the glaciers by +/- 0.05 to 0.1. The meteorological values used were gained from the monthly averages between 1975 to 1990 for the weather station at Ny-Ålesund (Hanssen-Bauer *et al.* 1990).

These tests were carried out using the Lovénbreen area distribution. Because we are interested in the specific mass balance profiles, and since the same meteorological data will be applied to each glacier, it is irrelevant which area distribution was used as this would only affect the total specific balance. The results of these tests are presented in appendix Three (figures A3.1 to A3.12) and Chapter Six (table 6.1). The version of the model used for these exercises excluded the latent heat component in the energy balance equation, and did not include the altitudinal precipitation correction.

5.3 Comparisons between the different methods of applying the meteorological data

The first objective of this project was to compare the results obtained from the Oerlemans (1991) method of applying meteorological data (Section 2.4.2) to the model and when measured weather data was used. The procedure was to carry out modelling using the Oerlemans (1991) method of defining the meteorological parameters (Section 3.4.2). These results were then compared with those obtained when one of the simplified parameters was replaced by its measured equivalent. For example, daily temperature calculated using the sinusoidal function was replaced with recorded temperature data from Ny-Ålesund, with the other values remaining as per Oerlemans (1991). This was done for each of the meteorological parameters; temperature, precipitation, humidity and cloudiness, and each of the balance years examined.

The results are presented in Chapter Six (figures 6.1 and 6.2). Modelling was carried out including and excluding the latent heat flux. No corrections were carried out for precipitation variation with altitude.

5.4 Modelling the mass balance of the glaciers Lovénbreen and Brøggerbreen for present day conditions

The second objective of this project was to reproduce the measured mass balances of Lovénbreen and Brøggerbreen for present day climatic conditions. Measured meteorological data were usually used, and there were two main variations worked with:

1. Use of the model with and without the latent heat flux being included in the energy equation.
2. The use of an altitudinal correction equation for the precipitation.

The altitudinal correction was determined for three seasons; 1983/83, 1985/86 and 1986/87. This was defined by the use of snow accumulation diagrams in Liestøl (1986), Hagen and Liestøl (1987) and Hagen (1988). These diagrams represent the snow accumulation around mid-May when the accumulation measurements were made. The values of snow accumulation were normalized to the accumulation at sea level (taken to be the value at day 150) and had a second order polynomial equation fitted to them, resulting in the equation below:

$$S_a = S (C_o + C_1 h + C_2 h^2)$$

where S_a is the snow accumulation at altitude h, S is the snow fall at sea level or in this case, that recorded at Ny-Ålesund, and C_o, C_1 and C_2 are constants appropriate for the 2nd order polynomial being used. It is these three constants which would account for the interannual variation for each glacier. Figures 5.2 and 5.7 show the variation with altitude of the snow accumulation. These approximately agree with a comment made by Liestøl (1988, 1990) that at sea level there is snow depth of around 1 metre and at the tops of mountains of 3 metres. The accuracy of these graphs is limited because of the coarseness of the snow accumulation scale used on the

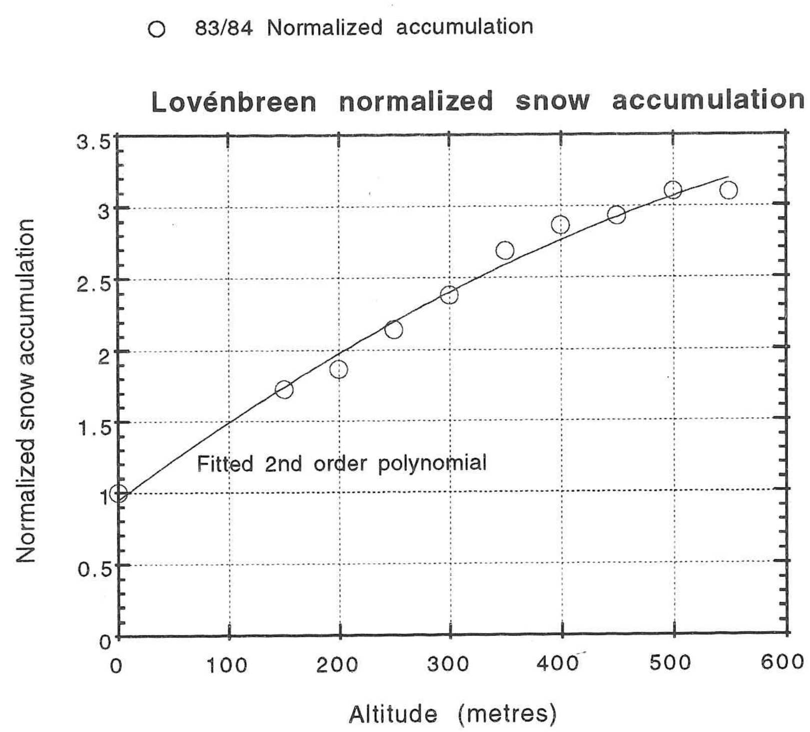


Figure 5.2 The snow accumulation with altitude for the glacier Lovénbreen for the season 1983/84 (Liestøl, 1986).

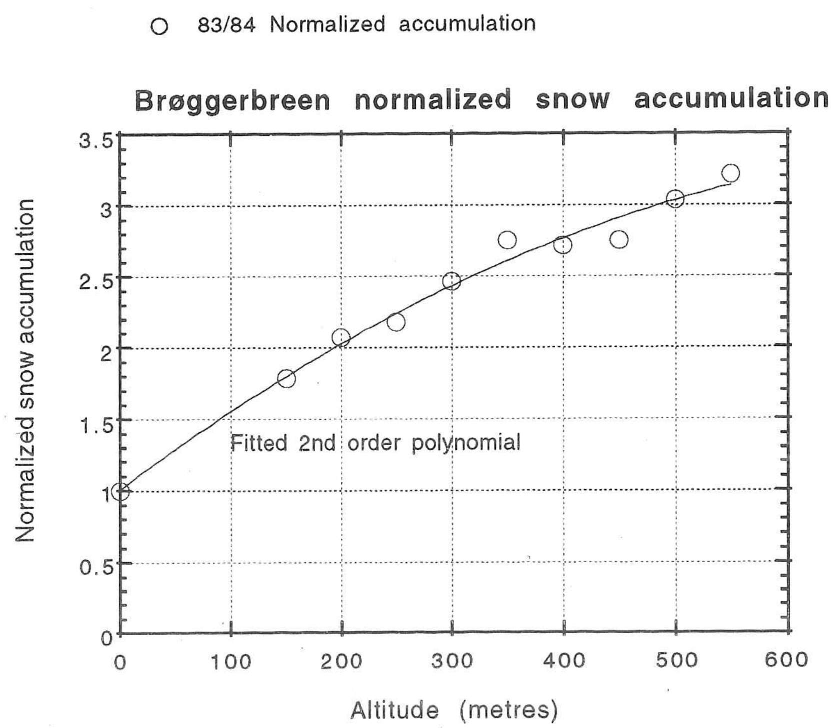


Figure 5.3 The snow accumulation with altitude for the glacier Brøggerbreen for the season 1983/84 (Liestøl, 1986).

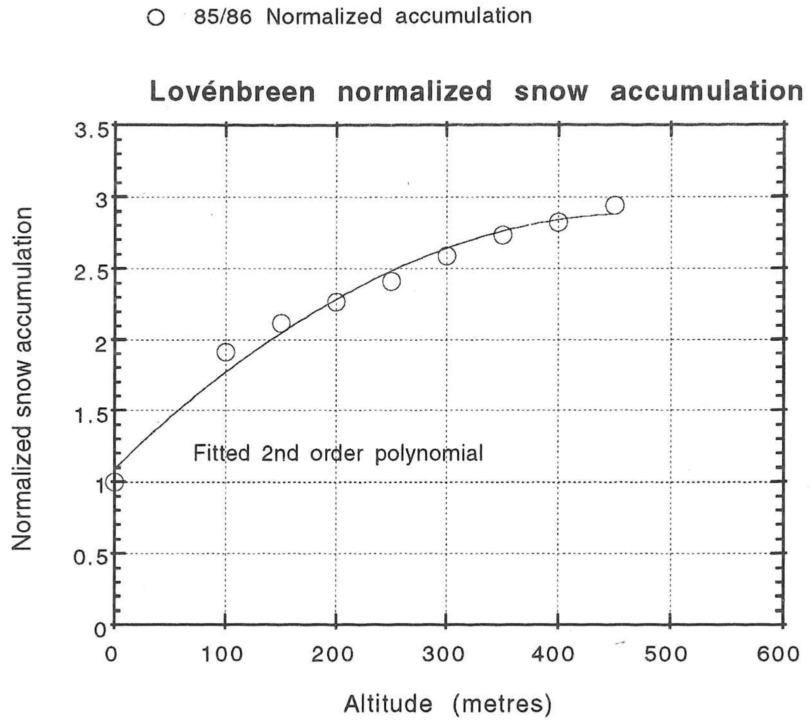


Figure 5.4 The snow accumulation with altitude for the glacier Lovénbreen for the season 1985/86 (Hagen and Liestøl, 1987).

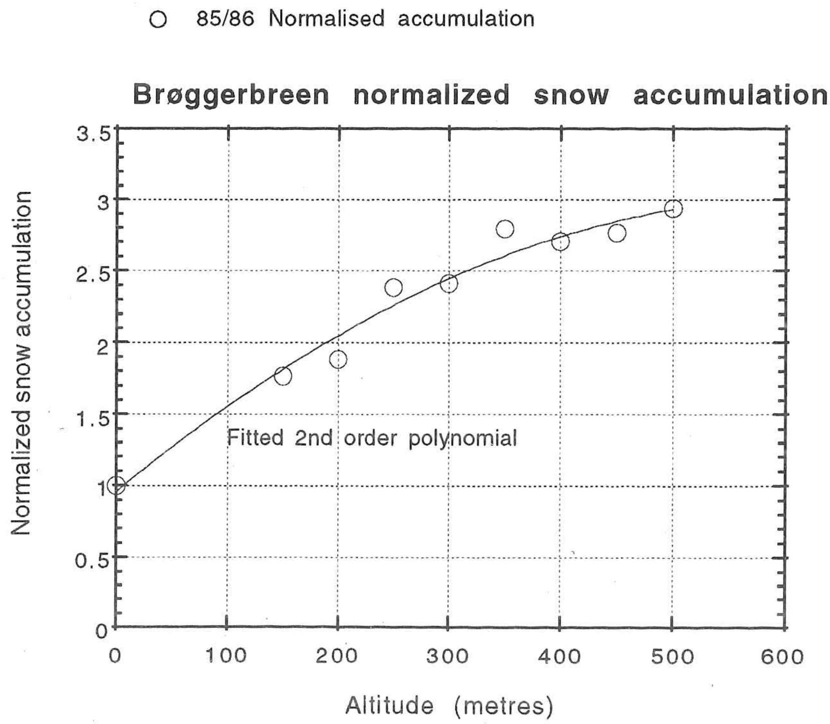


Figure 5.5 The snow accumulation with altitude for the glacier Brøggerbreen for the season 1985/86 (Hagen and Liestøl, 1987).

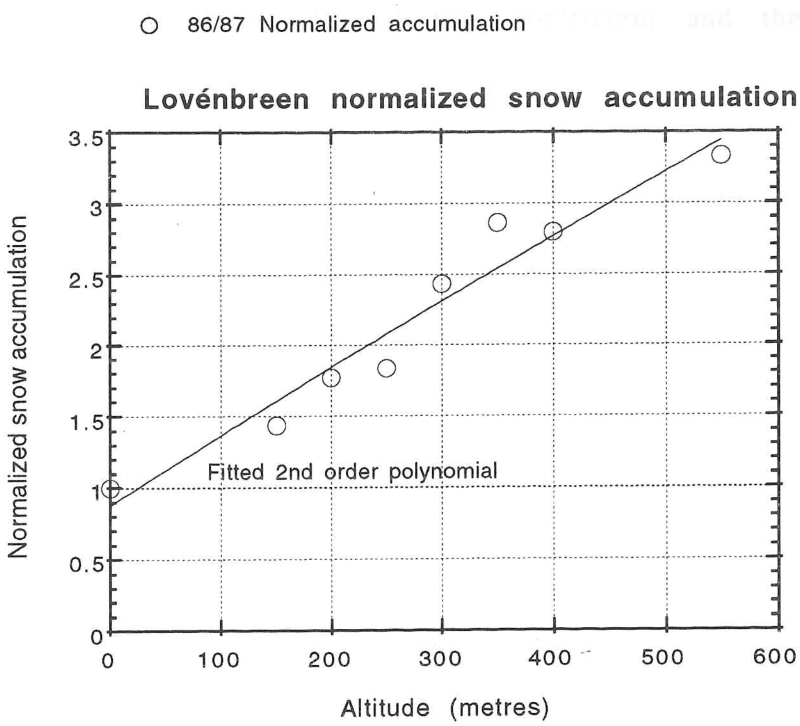


Figure 5.6 The snow accumulation with altitude for the glacier Lovénbreen for the season 1986/87 (Hagen, 1988).

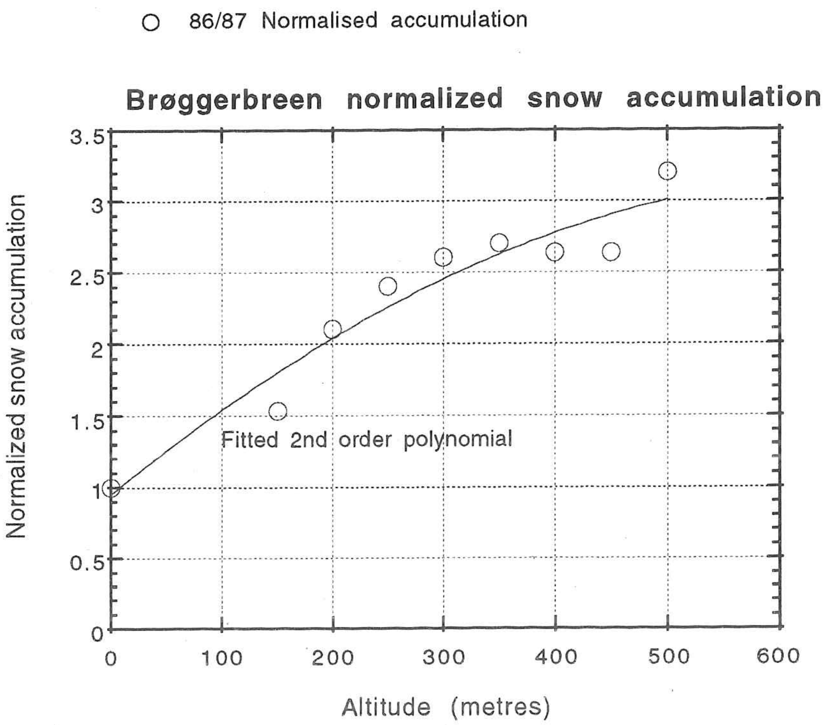


Figure 5.7 The snow accumulation with altitude for the glacier Brøggerbreen for the season 1986/87 (Hagen, 1988).

diagrams. Few efforts were made to *fine tune* the model by modifying parameters such as the turbulent heat flux coefficient and the glacier's surface albedo.

5.5 Modelling the mass balance of Spitsbergen glaciers for future and past conditions

The final objective of this project was to model the mass balances of these glaciers for the proposed warmer conditions due to greenhouse gases, and for the cooler conditions of the Little Ice Age.

Modelling was carried out for one year and one glacier, Brøggerbreen for the 1985/86 season. The reason for choosing that season is that the mass balance recorded that year was close to the average, and it was the example most closely modelled in the previous objective. The versions of the model used were those which excluded and included the latent heat component and included the altitudinal variation, as these gave the best results in the previous exercises on simulating present conditions.

The procedure in these exercises was to choose two increases in temperature and precipitation which might be expected to result from the greenhouse effect. These values are the upper and lower values obtained from Koster (1991) which are those expected from a doubling in the CO_2 concentration in the atmosphere. The values were 4° and 5°C for temperature, and precipitation was increased by 10 and 50%. The model exercises carried out were:

Annual mean temperature + 4°C
Annual mean temperature + 4°C and precipitation + 10%
Annual mean temperature + 4°C and precipitation + 50%
Annual mean temperature + 5°C
Annual mean temperature + 5°C and precipitation + 10%
Annual mean temperature + 5°C and precipitation + 50%

Another test carried out was to investigate how a change in the area's intra-annual variability would affect the resulting mass balances. This was studied by applying a seven-point smoothing function to the temperature data. This still gave the general trend of the annual temperature of Ny-Ålesund but did not have the daily variation. This was applied to the present day climate of Lovénbreen, excluding the latent heat component and the altitudinal correction for the precipitation.

The second part of this objective was to model the mass balance of these glaciers for the conditions of the Little Ice Age. This is difficult because of the uncertainty involved in actually knowing what the conditions at that time were. Simões (1991) determined from ice core data that the temperature increase for the 20th century average against that for the previous three centuries was between 2.3 and 3.3°C, for the ice caps Vestfonna and Austfonna on the island of Nordaustlandet, or 1.5 to 2.2°C for the Svalbard archipelago. Hence, the annual mean temperatures modelled were 2 and 3°C lower than the present day. The 1985/86 season for Brøggerbreen was again used. The model was run including and excluding the latent heat flux.

Because of time constraints on this project, a more complete range of modelling exercises could not be preformed. Other possible options which may yield useful information will be mentioned in Chapter Seven.

Chapter Six
Mass balance modelling of Spitsbergen glaciers: Results and interpretation

This chapter will present the results and interpretations of the modelling experiments carried out as described in Chapter Five. These tests were a series of climatic sensitivity studies to show how perturbations in different parameters affected the results, and the three objectives of this project. These objectives were 1) examining how different ways of applying meteorological data to the model affected the results, 2) modelling the mass balance of two Spitsbergen glaciers, Lovénbreen and Brøggerbreen, for present day climatic conditions and 3) modelling the mass balances of these glaciers for future (warmer) and past (cooler) climates. The final two scenarios will represent the proposed conditions due to greenhouse warming and the Little Ice Age.

6.1 Sensitivity studies

The sensitivity studies were carried out to determine what effect varying each of a series of parameters had on the calculated mass balance. The results of these exercises are summarized in table 6.1, and figures A3.1 to A3.12 in Appendix Three. The parameters varied were the annual mean temperature, annual precipitation, mean annual humidity and cloudiness, cloud base height and the glacier's albedo.

6.1.1 Results

It is obvious from the figures presented in Appendix Three that the effect on the mass balance due to the perturbations in annual mean temperature (figures A3.1 and A3.2), while modest (+/- 1 and 2^{o}C), had a greater effect than the variations of the other parameters by almost an order of magnitude. The next most influential parameter was the surface albedo (figures A3.11 and A3.12), followed by annual precipitation (figures A3.3 and A3.4), cloudiness (figures A3.7 and A3.8) and humidity (A3.5 and A3.6). The parameter which had the least effect was cloud base altitude (figures A3.9

and A3.10). These points are further described by table 6.1 which presents the differences between the reference calculations and the results from the climatic perturbations.

6.1.2 Interpretation

For all of the tests, the largest differences between the reference profile and those resulting from the perturbations in the various parameters were in the lower part of the glacier, especially below the equilibrium line altitude.

The two profiles representing differences obtained from decreasing the mean annual temperature (figure 3.2) coincide for the upper part of the glacier, especially noticeable in the profile figure (figure A3.1). This is because the decrease in both instances was sufficient to ensure there was no melting of snow in the upper part of the glacier. The dominant reason for the major changes in specific balance for the temperature variations result from changes in the length of the ablation season. The change in the length of the season going from the reference run to the 1°C increase is approximately 20 days.

The changes due to the variations in precipitation (figures A3.3 and A3.4) differed from those found by Oerlemans and Hoogendoorn (1989). In that work, it was found that variations in the mass balance profiles were almost independent of altitude. This was not the case for the examples presented in this work, with the change in precipitation being greatest at the lower part of the glacier. Like Oerlemans and Hoogendoorn (1989), the change in albedo around the equilibrium line was greatest, and decreased with altitude. The effect of the precipitation was less at the upper most parts of the glacier, since precipitation at these altitudes was usually snow and the change in balance at these heights was merely the change in the amount of precipitation.

The humidity tests (figures A3.5 and A3.6) should be treated with suspicion because of the fact that one of the components which is directly influenced by humidity, the latent heat flux, was not included in these calculations. The only part of the formulation affected by this variation was the incoming longwave radiation.

Change in equilibrium altitude (metres)		Change in specific balance (mm yr^{-1} of water equivalent)
Temperature tests		
Test 1 +1°C	129	-775
Test 2 +2°C	255	-1592
Test 3 -1°C	-126	666
Test 4 -2°C	-253	1141
Precipitation tests		
Test 5 +10%	-39	152
Test 6 +20%	-56	290
Test 7 -10%	33	-162
Test 8 -20%	64	-331
Relative humidity		
Test 9 +10%	22	-129
Test 10 +20%	41	-247
Test 11 -10%	-18	111
Test 12 -20%	-39	226
Cloudiness		
Test 13 +10%	-17	149
Test 14 +20%	-40	307
Test 15 -10%	14	-140
Test 16 -20%	29	-276
Cloud base altitude		
Test 17 +10%	-5	29
Test 18 +20%	-10	59
Test 19 -10%	5	-30
Test 20 -20%	10	-61
Albedo		
Test 21 +10%	-47	310
Test 22 +20%	-97	586
Test 23 -10%	47	-343
Test 24 -20%	93	-711

Reference test results: Equilibrium line altitude 540 m
Specific mass balance -1146 mm yr^{-1}

Table 6.1 How the equilibrium line altitude and specific mass balance values vary for the sensitivity tests. The reference test used as climatic input meteorological parameters defined by the method of Oerlemans (1991) and found from the averaged climatic values measured at Ny-Ålesund between 1975 to 1990 (Hanssen-Bauer, 1990).

Change in equilibrium (metres)	altitude	Change in specific balance (mm yr^{-1} of water equivalent)
Temperature tests		
Test 1 +1°C	129	-775
Test 2 +2°C	255	-1592
Test 3 -1°C	-126	666
Test 4 -2°C	-253	1141
Precipitation tests		
Test 5 +10%	-39	152
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Test 8 -20%	64	-331
Relative humidity		
Test 9 +10%	22	-129
Test 10 +20%	41	-247
Test 11 -10%	-18	111
Test 12 -20%	-39	226
Cloudiness		
Test 13 +10%	-17	149
Test 14 +20%	-40	307
Test 15 -10%	14	-140
Test 16 -20%	29	-276
Cloud base altitude		
Test 17 +10%	-5	29
Test 18 +20%	-10	59
Test 19 -10%	5	-30
Test 20 -20%	10	-61
Albedo		
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Test 24 -20%	93	-711

Reference test results: Equilibrium line altitude 540 m
Specific mass balance -1146 mm yr^{-1}

Table 6.1 How the equilibrium line altitude and specific mass balance values vary for the sensitivity tests. The reference test used as climatic input meteorological parameters defined by the method of Oerlemans (1991) and found from the averaged climatic values measured at Ny-Ålesund between 1975 to 1990 (Hanssen-Bauer, 1990).

Changes in cloudiness had a relatively large effect (figures A3.7 and A3.8), again a function of the incoming longwave radiation and also the shortwave. There was minimal difference around the upper parts of the glacier as found by Oerlemans and Hoogendoorn (1989). Here the changes in the longwave and shortwave contributions will cancel each other out. However at the lower altitudes the increase in the shortwave will be greater than the decrease in longwave, producing a more negative balance for lower cloudiness, with the opposite for increased cloudiness.

The height of the cloud cover had the least effect on the results (figure 3.9 and 3.10). This is fortunate as the height of the cloud base is not well known. Its contribution was the variation in the incoming longwave energy component. A higher cloud cover will be at a lower temperature, hence have a smaller longwave contribution, with the opposite effect for lower clouds.

Altering the albedo had the second largest effect (figure A3.11 and A3.12). The concern with this parameter is that the magnitude of the variations, +/- 0.05 to 0.10, are quite small and, as commented upon by Oerlemans and Hoogendoorn (1989), may be of the order of the uncertainty in measurements of this parameter.

6.2 Results from the different methods of applying the meteorological data

6.2.1 Results

One objective of this project was to compare the ways in which meteorological data may be applied to the model. As described in earlier parts of this thesis (Section 3.4.2, 5.2), Oerlemans (1991 and pers. comm.) made a number of assumptions with regards to this aspect of the model's operation. The assumption was that the simplified data input used by Oerlemans (1991) would have little effect on the results relative to those which would be gained from the use of real data. It was this assumption that was tested. Another option tested was the inclusion or exclusion of the latent heat component of the energy equation. The comparisons between the various methods are shown in figures 6.1 and 6.2.

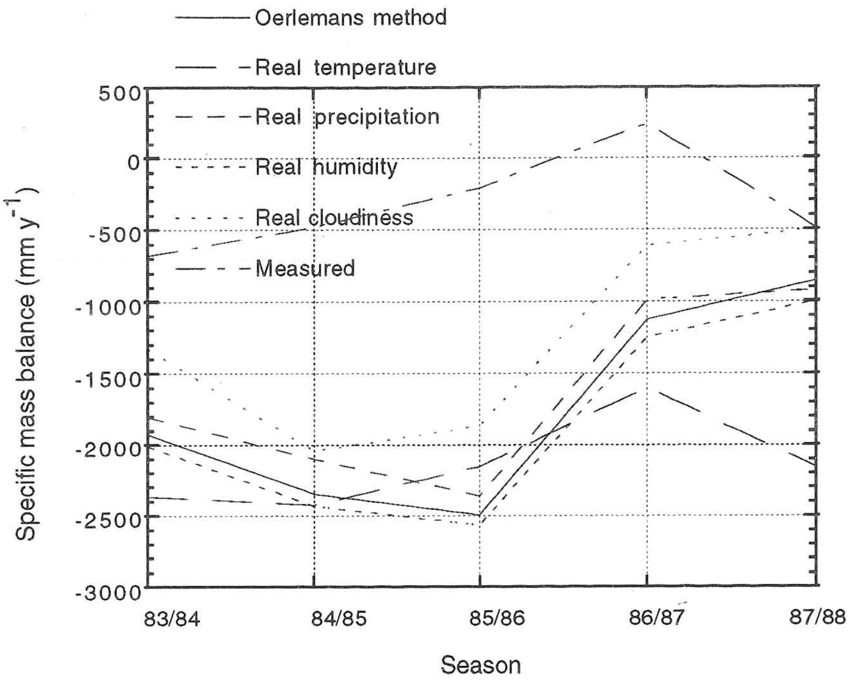

Figure 6.1 The different results gained from using the Oerlemans (1991) method of applying meteorological data to the mass balance model, and the use of measured climatic data. The latent heat flux has not been included in the energy equation.

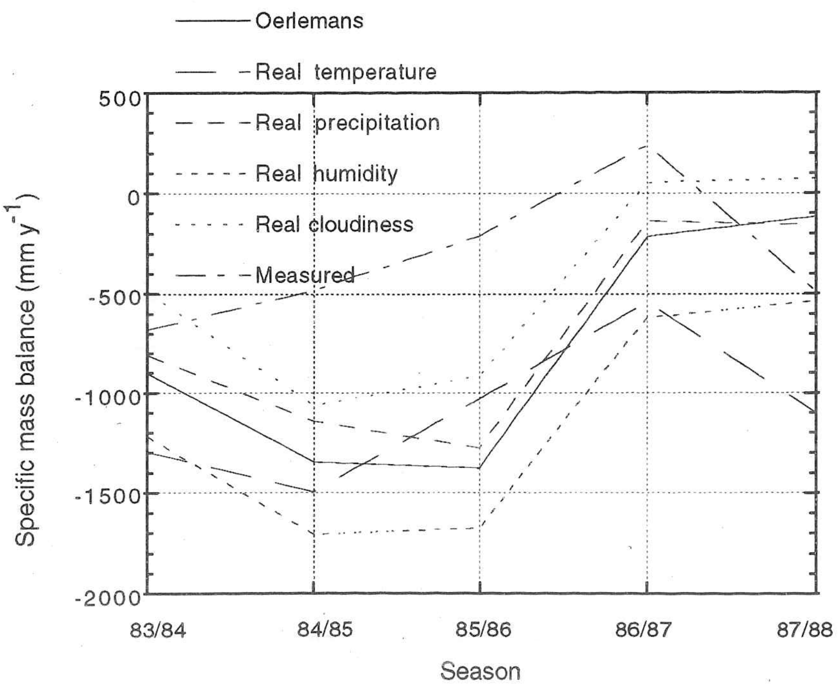

Figure 6.2 The different results gained from using the Oerlemans (1991) method of applying measured meteorological data to the mass balance model, and the use of measured data. The latent heat flux has been included in the energy equation.

6.2.2 Interpretation

From figures 6.1 and 6.2, it is seen that substituting real precipitation had little effect on the results for both including and excluding the latent heat flux. This is consistent with Oerlemans (in press) where several different arrangements of precipitation distribution were tried, with little difference noted. The factor which had the greatest effect was the inclusion of measured temperature values, expected in light of our previous discussion concerning the sensitivity tests (section 6.1). Substituting for measured cloudiness had the next greatest effect, followed by humidity which showed a greater difference when the latent heat was included.

It is obvious from these graphs (figures 6.1 and 6.2) that the model, at least in these instances, has failed on two accounts. The first is that the resulting mass balances are incorrect when compared with the measured mass balances values. The second is that the general trend of the measured mass balances are poorly represented by the modelled results, except for the option where the real temperature data have been used.

The importance of the tests discussed in this section is not whether each option for the application of meteorological data could reproduce the mass balance behaviour of the glacier being modelled, but if the final results were significantly affected by how the initial data is selected. It is found that this is the case for temperature and cloudiness.

6.3 Modelling results for the glaciers Lovénbreen and Brøggerbreen for present day conditions

This objective is the most important, as the next, which deals with the modelling of future and past conditions, depends upon the model being able to reproduce the mass balances recorded from contemporary field surveys. The results of these exercises are summarised in figures 6.3 and 6.4, which show the resultant modelled mass balances for the glaciers Lovénbreen and Brøggerbreen. The options tested were:

1. Oerlemans' (1991 and pers. comm) method of applying the meteorological data

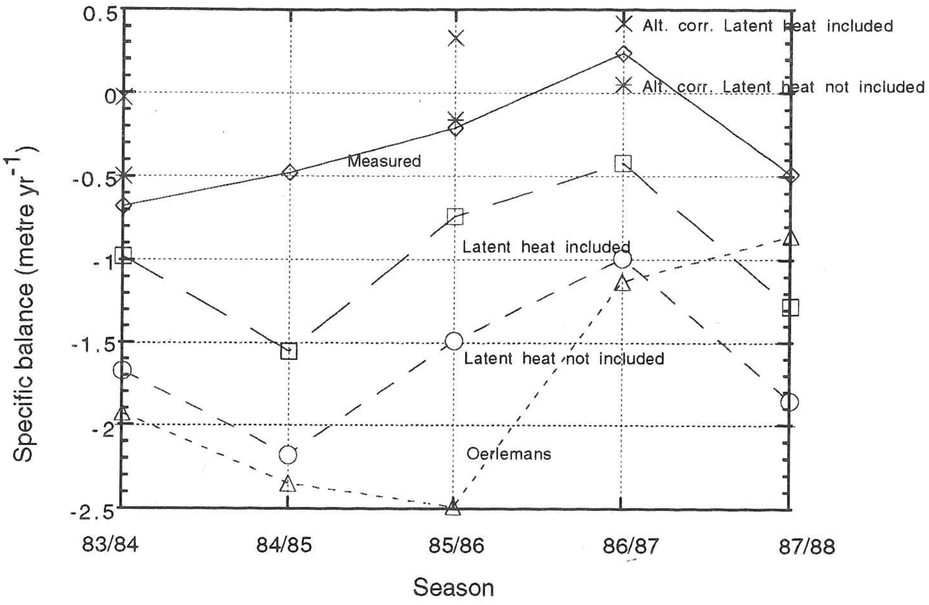

Figure 6.3 The mass balance modelling results for the glacier Lovénbreen.

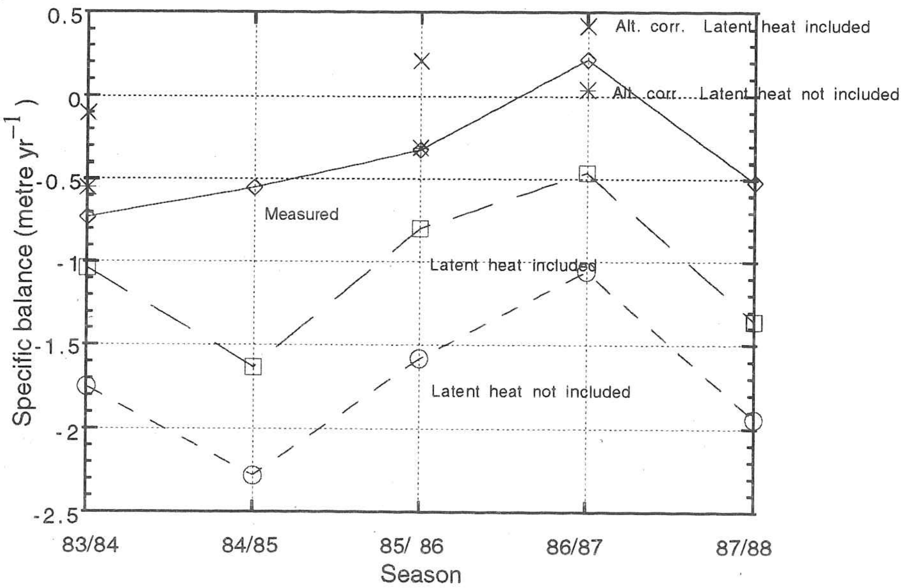

Figure 6.4 The mass balance modelling results for the glacier Brøggerbreen.

2. The use of measured meteorological data
3. The use of an altitude correction function for the precipitation

In each one of these, modelling was carried out with and without the latent heat component being included.

6.3.1 Results

It is apparent from figures 6.3 and 6.4 that the results using the altitudinal variation and not including the latent heat flux were the best. Unfortunately, the information was not available to produce the precipitation/altitude function for snow accumulation variation for the seasons 1984/85 and 1987/88. The other results (excepting those gained from the Oerlemans (1991) method of applying the meteorological data) followed the general trend of the seasons, that is there was an increase in balance for the seasons 1983/84 to 1986/87, with the exception of 1984/85, with a decrease for 1987/88. The use of raw data proved to be superior to the simplified meteorological input of Oerlemans (1991) both in obtaining the year to year trend in the balances and in the actual values.

6.3.2 Interpretation

As mentioned in Section 6.3.1, the use of the altitudinal correction for precipitation gave the best results. This would be expected as precipitation does vary with altitude (Liestøl, 1988, 1990) and this should be accounted for in any mass balance modelling operation. The best result of these was for Brøggerbreen and the 1985/86 season where the modelled answer was -31 cm water equivalent, compared with -32 cm measured. The exclusion of the latent heat seemed also to be the better option, however reasons for this are not apparent.

Little tuning was carried out on the model to produce better results. One experiment done on the Lovénbreen glacier for 1983/84 introduced a +0.1 increase to the background albedo. This did give a mass balance result closer to that measured (-70 cm water equivalent modelled compared with -68 cm measured), however there was no justification for introducing this increase, so this approach was abandoned. The danger with such an approach is that the situation could be reached where the model is simply mimicking the glacier, and that it is useless for other glaciers under different conditions.

6.4 Modelling results for Spitsbergen glaciers for future and past conditions

This part of the project depended heavily upon the success of the previous part (section 6.3). Naturally, if the mass balance situation today could not be reproduced, then there is little point in pursuing further modelling exercises.

6.4.1 Modelling for warmer climates

The first exercise was for the proposed future conditions resulting from greenhouse warming. The situation was for a doubling of the CO_2 content in the atmosphere. Only variations in the annual mean temperature and precipitation were modelled, using values from Koster (1991). Because the best result obtained in the previous section was for the glacier Brøggerbreen in the season 1985/86 using the altitudinal correction for precipitation, it was this glacier/season combination and model that was worked with in these exercises. The modelling was carried out including and excluding the latent heat flux.

The results of these exercises are presented in table 6.2. For the option where the latent heat was not included, despite a 50% increase in precipitation, a 4^{o}C increase resulted in only the upper most region of the glacier maintaining a positive mass balance. When this is raised to 5^{o}C, the entire glacier experiences a negative balance. When the latent heat is included, the upper part of the glacier maintains a positive balance for the 4^{o}C increase, and only does so for the 5^{o}C increase when there is a 50% in the precipitation. These variations would in reality be offset by the increased cloud formation which would accompany these changes, however as shown by the sensitivity studies (section 6.1, figures A3.7 and A3.8), this has the greatest effect at the lower altitudes with little effect at the higher points, and the magnitude of the cloudiness effects are very much less than those due to temperature (compare figures A3.1 and A3.2 with A3.7 and A3.8).

	Equilibrium altitude	Specific mass balance
+4°C	799 metres	-3.32 m
+4°C + 10% precipitation	750 metres	-3.13 m
+4°C + 50% precipitation	622 metres	-2.45 m
+5°C	1000 metres	-4.24 m
+5°C + 10% precipitation	934 metre	-4.06 m
+5°C + 50% precipitation	761 metres	-3.39 m
Present day	339 metres	-0.31 m

Resultant equilibrium altitudes and mass balances for warmer conditions when the latent heat has not been included.

+4°C	637 metres	-2.18 m
+4°C + 10% precipitation	598 metres	-1.99 m
+4°C + 50% precipitation	496 metres	-1.38 m
+5°C	772 metres	-2.99 m
+5°C + 10% precipitation	726 metres	-2.80 m
+5°C + 50% precipitation	609 metres	-2.16 m
Present day	195 metres	-0.42 m

Resultant equilibrium altitudes and mass balances for warmer conditions when the latent heat has been included.

Table 6.2 Results of the modelling experiments for the proposed increases in annual mean temperature and precipitation.

6.4.2 Modelling for cooler climates

For the modelled conditions representing the cooler Little Ice Age climate, the results are presented in table 6.3. There is now the situation of a large positive budget for much of the glacier's surface. For the situation where the latent heat is included, the result of a 3°C decrease is that the equilibrium line altitude extends to below the position of the present glacier front, around 50 metres (Norsk Polarinstitutt, 1979), meaning that it has a positive budget over the entire present extent of the glacier. The exclusion of the latent heat flux will still causes a large positive budget with a much lower equilibrium line altitude.

	Equilibrium altitude	**Specific mass balance**
-2°C	178 metres	0.52 m
-3°C	To sea level	1.03 m
Present day	339 metres	-0.31 m

Resultant equilibrium altitudes and mass balances for warmer conditions when the latent heat has not been included.

-2°C	104 metres	0.79 m
-3°C	To sea level	1.03 m
Present day	195 metres	-0.42 m

Resultant equilibrium altitudes and mass balances for warmer conditions when the latent heat has been included.

Table 6.3 Results of the modelling experiments for the proposed decreases in annual mean temperature.

6.4.3 Modelling a change in climate variability

One other modelling exercise carried out was for the case where the climatic variability of an area changed. This was carried out by the use of a 7-point moving point smoothing function. The results are presented in figure 6.5. This was tested for one situation, the glacier Lovénbreen over the five balance years. The latent heat flux was not included, nor was the altitudinal variation in the precipitation. The parameter altered was the daily temperature. The result is a less negative balance for all of the seasons.

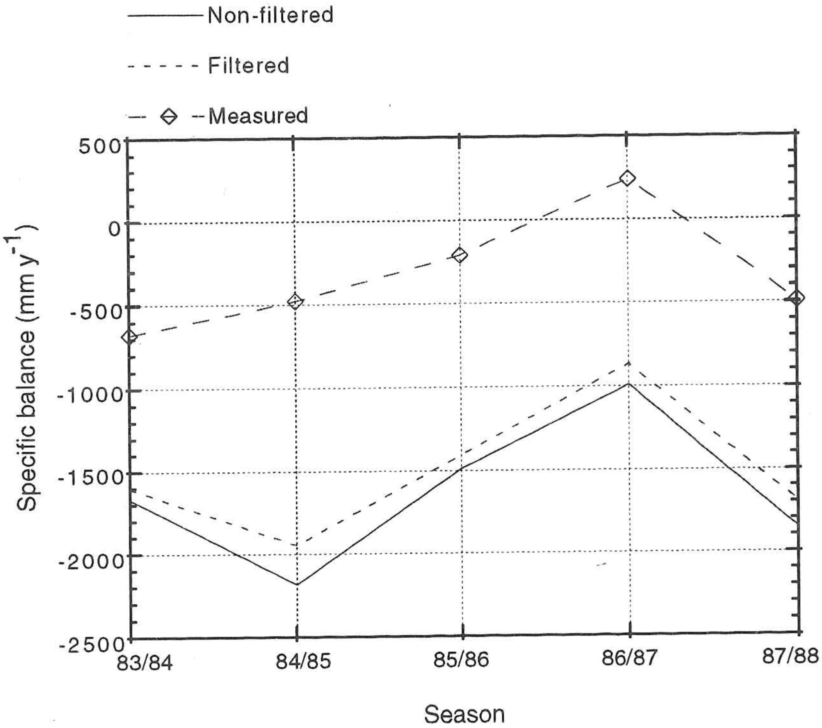

Figure 6.5 The mass balance modelling results for the glacier Lovénbreen when comparing the use of raw temperature data and that which has been smooth, simulating a decrease in climatic variability.

Chapter Seven
Conclusions

Chapter One of this thesis introduced the glaciological problem which was to be treated, the mass balance modelling of two Spitsbergen glaciers under different climatic conditions. The defined objectives were:

1. Comparing the mass balance results using different treatments of the meteorological data.
2. Modelling the mass balance of the glaciers for present day conditions
3. Modelling the mass balance of the glaciers for future (warmer) and past (cooler) conditions.

Chapter Two was concerned with the processes which influence the annual mass balance of a glacier. The methods utilised to determine glacier mass balances and the model used in this project were described in chapter Three. Chapter Four described the study area, the north-west of Spitsbergen. Chapter Five dealt with the procedures for the modelling exercises and Chapter Six discussed the results. This chapter will provide a summary of these results and make some suggestions for future work.

7.1 Summary of the results

The sensitivity studies proved to be useful in that they assisted in identifying which parameters the model was most receptive too, and which one had little impact on the final mass balance results. The temperature and surface albedo of a glacier were found to be the most influential with the cloud height being of little importance (table 6.1).

The manner in which the meteorological data are defined for the model has significant importance. This was evident from the variation in the results gained when each of the averaged meteorological values was substituted for its measured equivalent (figures 6.1 and 6.2).

There is reason for some confidence in the use of this model to reproduce the measured mass balances of Spitsbergen glaciers when the option incorporating an altitudinal variation in precipitation and excluding the latent heat flux is used (figure 6.1). The inclusion of the precipitation variation with altitude is a more realistic representation of the situation involving the glaciers Lovénbreen and Brøggerbreen. The validity of excluding the latent heat flux is uncertain, and requires further investigation.

Past and future climates modelled show that for a 4 to 5°C increase in temperature and a 10 to 50% increase in precipitation, the glacier Brøggerbreen will be at a negative balance for much of its present extent when imposing this temperature and precipitation increase on the 1985/86 season meteorological values. For example, a +5°C increase in the annual mean temperature, with the 50% increase in precipitation when not including the latent heat flux will result in the glacier being at a negative balance for its entire extent (table 6.2). This is apparent by the equilibrium line altitude being at 761 metres for these conditions, as opposed to 339 metres for the present day situation (as calculated by the model).

The situation for the proposed Little Ice Age conditions is the reverse. A much greater proportion of the glacier will be at a positive balance while for a decrease of 3°C, the region of the glacier at positive balance will extend to below the present lowest point of the glaciers. Even for a 2°C decrease, this will result in the equilibrium line being at 178 metres when the latent heat has been excluded, compared with 339 metres (table 6.3).

7.2 Proposals for future work

The time constraints on this project meant that a more comprehensive and complete series of modelling exercises could not be carried out. Therefore, the first suggestion for future work is to remedy this by carrying out a greater number of modelling tests. These additional tests will cover such aspects as:

1. For the present day situation, expanding the tests from the two glaciers studied during this project to others in the vicinity, and then to other areas of Svalbard. This will require the use of data from the other weather

stations located on the archipelago. Also, extending the range of balance years modelled, provided the data exists, would be done at the same time.

2. For modelling warmer climates, the testing of different predictions for changes in temperature, precipitation, cloudiness, daily and seasonal variability, resulting from a range of climate and atmospheric models.

3. For past cooler climates, again different scenarios for the different periods would need be tested. Data for these would be acquired from sources such as ice core studies.

4. Modelling the present day situation of a larger number of glaciers from different climatic regions around the world. This will allow the evaluation of what components of the energy budget are important for each region, and to develop better ways of correcting available meteorological data for the local conditions.

The second suggestion concerns additional work on the treatment of the input data. Because of the expense and effort involved, year-round meteorological observations are impractical for the entire surface area of a glacier. One parameter which could be better defined is the surface albedo. The use of digital satellite imagery would come into play here. This line of research is at present being pursued at the Scott Polar Research Institute.

The model itself also needs improvement, the areas with scope for this are the same as those suggested by Oerlemans (1991). These include:

1. The surface albedo
2. The treatment of meltwater and formation of superimposed ice
3. The turbulent energy fluxes
4. The radiation balance and the influence of clouds

As there were tests conducted with and without factors such as the latent heat flux, the relative importance of such process in a study area would need to be determined. Factors such as the influence of volcanic dust in the atmosphere, which may be an important factor (Oerlemans, 1988), could also be incorporated.

The final suggestion concerns the comment made at the start of this thesis, that this model may form the subsection of a larger, more comprehensive model. This larger model would account for such features as:

1. The dynamics of the ice mass being examined
2. The heat budget of the ice mass, both on its surface and internally, to better account for processes such as surface and basal melting
3. The ability at any point to introduce an external event such as a volcanic eruption, and to see what the response of the ice mass is
4. Coupling between the atmosphere and the ice mass, which will incorporate the assorted feedback effects

Such a model would be time dependent and would also include such features as the geometry of the ice mass to be examined. A model which covers some of these points is that of Oerlemans (1988).

In conclusion, the model used in this project is a first step towards the comprehensive and accurate modelling of the behaviour of ice masses in response to climatic change. As techniques such as airborne and satellite remote sensing become more sophisticated, and as more data concerning past conditions comes forth, there will be ample opportunities to test and refine any new model. With the increasing power of computers, the ability to incorporate more parameters into a single model also becomes practical.

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Appendix One
The mass balance model

This appendix will present the mass balance modelling program used throughout this project. The version shown is that which includes the altitudinal correction for precipitation and excludes the latent heat flux part of the energy equation.

Examples of the input files are also shown. These are for defining the area distrbution of the glacier and the climate for a specific year.

This model was run on an Apple Macintosh computer. The code is in standard FORTRAN-77 with no machine dependent features to the knowledge of the author.

A1.1 The model program

```
      PROGRAM  MASS_BALANCE_MODEL
C-----------------------------------------------------------------------
C  Mass balance modelling program using the energy balance model of:
C       Oerlemans, J. (1991)
C       The Holocene, Volume 1,  number 1,  pages 40 to 48
C
C  Program written by:  J. Oerlemans
C                       Institute for Marine and Atmospheric Research
C                       University of Utrecht, Utrecht, The Netherlands
C  Modified by:         K. Fleming
C                       Scott Polar Research Institute
C                       University of Cambridge, Cambridge, England
C-----------------------------------------------------------------------
C  Initialisation of parameters and arrays.

      IMPLICIT NONE

      INTEGER  i,     j,     k,     l,    x,     y,     z,    nxx,     nx,
     :         nday, ndaytot, ndaymax,  klm,  numrun, daysrt,  day

      REAL     longin,  long,     prectot,  specbal,  totarea,  ella,
     :         xlapse,  twpi,     halfpi,   a,        b,        enerb1,
     :         enerb2,  tc,       trest,    ttot,     snacc,    clcontr,
     :         emmis,   press,    eas,      ea,       fcl,      tcloud,
     :         sigma,   rad,      precip,   temp,     tempy,    cldhei

      REAL     alti,     sola,     senflux,  albedo,  albfin,   snow,
     :         snowmax,  precalt,  exch,     albsnow, tice,     accum,
     :         enerbal,  area,     massbal,  latflux, precal,
     :         prec0,    prec1,    prec2

      REAL     xinsol,   precipt,  temper,   relhum,   xn

*     CHARACTER*20     GLACIER_DATA,    METELOG_DATA,    RESULT1,
     :                 RESULT2

      DIMENSION alti(50),     sola(50),     senflux(50),  albedo(50),
     :          albfin(50),   snow(50),     snowmax(50),  precalt(50),
     :          exch(50),     albsnow(50),  tice(50),     accum(50),
     :          enerbal(50),  area(50),     massbal(50),  latflux(50),
     :          long(50)

      DIMENSION xinsol(365),  precipt(365),  temper(365),  relhum(365),
     :          xn(365)

C-----------------------------------------------------------------------
C    massbal: accumulated mass budget from day daysrt -->Bn            |
C    accum:   mass accumulation from day daysrt                        |
C    snow:   snow budget from daysrt                                   |
C    precalt: altitudinal distr. of precip / unit: 1000 mm             |
C    exch:   exchange coefficient for turbulent fluxes                 |
C-----------------------------------------------------------------------
C  The required data files are named, opened and/or created.
C  There is the ability to correct errors in file name input.

      WRITE(*, 300)
  100 CONTINUE
```

```
      WRITE(*, 400)
      READ(*, '(A20)')   GLACIER_DATA
      OPEN(UNIT=10, FILE=GLACIER_DATA, STATUS='old', ERR = 150)
      WRITE(*, 500)
      READ(*, '(A20)')   METELOG_DATA
      OPEN(UNIT=20, FILE=METELOG_DATA, STATUS='old', ERR = 150)
      WRITE(*, 600)
      READ(*, '(A20)')   RESULT1
      OPEN(UNIT=40, FILE=RESULT1,      STATUS='new', ERR = 150)
      WRITE(*, 700)
      READ(*, '(A20)')   RESULT2
      OPEN(UNIT=50, FILE=RESULT2,      STATUS='new', ERR = 150)

      GOTO 200

  150 CONTINUE
      WRITE(*, 800)
      READ(*, *)
      GOTO 100

  200 CONTINUE
      OPEN(UNIT=30, FORM='formatted', FILE='INSO79N', STATUS='old')
      OPEN(UNIT=70, FILE='ENERGY',   STATUS='new')


  300 FORMAT(10X, ' MASS BALANCE MODELLING PROGRAM' /)

  400 FORMAT('Please enter glacier data file name:             ' $)

  500 FORMAT('Please enter meteorological data file name:      ' $)

  600 FORMAT('Please enter output profile data file name:      ' $)

  700 FORMAT('Please enter output parameter file name:         ' $)

  800 FORMAT(// 5X, ' SORRY, error in the file name '
     :         / 5X, '     Please press RETURN ' $)

C-----------------------------------------------------------------------
C  The required data from the files GLACIER_DATA, METELOG_DATA and
C  INS079N are read.

      READ(10, *)  totarea,  nx

      DO 900 i = 1, nx
             READ(10, *)  area(i),  alti(i)
  900 CONTINUE

      WRITE(50, 1000) (alti(i), i = 1, nx, 2)

 1000 FORMAT('Altitude.', 8X, 40F7.0)

      READ(20, *) ella, xlapse, cldhei, daysrt, prec0, prec1, prec2
      tempy   =  0.
      j       =  daysrt

      DO 1100 i = 1, 365
            READ(20, *)  day,  temper(j), precipt(j), relhum(j),
     :                   xn(j)
            tempy  =  tempy + temper(j)
            j = j + 1
            IF(j.EQ.366)  THEN
                 j = 1
```

```
          ENDIF
          READ(30, *)  xinsol(i)
 1100 CONTINUE

      tempy  =  tempy/365.

C-----------------------------------------------------------------------
C  Initialisation of the various parameters.  The main loop runs for as
C  long as is required for a stable solution to arise.

      nxx     = nx - 1
      rad     = 0.017453293
      twpi    = 6.2831853
      halfpi  = 1.5707963
      sigma   = 0.0000000567

      WRITE(*, 1500)
      READ(*, *)  numrun
      WRITE(*, *)  ' '

 1500 FORMAT(/ 'How many runs do you wish to do?   '$)

      DO 2100 klm = 1, numrun
            nday    = daysrt
            ndaytot =   1
            ndaymax = 365

C-----------------------------------------------------------------------
C  Loop sets parameters for each glacial segment.

            DO 1600 i = 1, nx
                  albfin(i) = 0.18*0.637*ATAN((alti(i)-ella+300.)
     :                              /200.) + 0.43
                  albsnow(i) = 0.75
                  exch(i)    = 7.
                  tice(i)    = tempy + alti(i)*xlapse
                  IF(tice(i).GT.0) THEN
                        tice(i)=0.
                  ENDIF
                  IF(klm.EQ.1) THEN
                        albedo(i)  = albfin(i)
                  ENDIF
                  massbal(i)  = 0.
                  snow(i)     = 0.
                  snowmax(i)  = 0.
                  accum(i)    = 0.
                  enerbal(i)  = 0.
                  precalt(i)  = 0.
 1600       CONTINUE

 1700       CONTINUE
            precip   =  precipt(nday)

C-----------------------------------------------------------------------
C  Cloud base temperature is calculated

            tcloud = 273.+ temper(nday) + (xlapse+0.00150 *
     :                    cos(twpi*(nday-26.)/365.)) * (cldhei)

            DO 1800 i=1, nx
                  precal   =  precip*(prec0 + prec1*alti(i) +
     :                          prec2*alti(i)**2)
                  temp = temper(nday) + (xlapse+0.00150 *
```

```
     :                          cos(twpi*(nday-26.)/365.)) *alti(i)
C------------------------------------------------------------------
C  Calculation of saturation vapour pressure over water

               eas = 610.8*exp(19.858*(1.-273.15/
     :                        (temp+273.15)))

C------------------------------------------------------------------
C  Calculation of vapour pressure from relative humidity with respect
C  to water

               ea   = relhum(nday) * eas

C------------------------------------------------------------------
C  Calculation of saturation vapour pressure over ice

               IF(enerbal(i).LT.0.)  THEN
                        eas = 610.8*exp(-22.47*(273.15/
     :                        (temp+273.15)-1.))
               ENDIF
               press=(1.-0.0001*alti(i))*1.e+5

               senflux(i)=exch(i)*temp
               latflux(i)=2488*0.622*exch(i)*(ea-eas)/press

               emmis=0.7+5.95e-7*ea*exp(1500./(temp+273.15))
     :                  -2.5e-5*alti(i)
               fcl=0.25
               clcontr=xn(nday)*fcl*tcloud**4
               longin=sigma*(clcontr+emmis*(temp+273.15)**4)
               long(i)=longin-315.6

C------------------------------------------------------------------
C  Insolation

               tc   = 1.-(0.41-0.000065*alti(i))*xn(nday)
     :                 -0.37*xn(nday)**2
               trest = (0.79+2.4E-5*alti(i))
               ttot  = tc*trest
               sola(i)  = xinsol(nday)*ttot

C------------------------------------------------------------------
C  Mass and energy budgets

               snacc=0.
               IF (temp.LT.2) THEN
                     snacc=precal
               ENDIF
               accum(i)  = accum(i)  + snacc
               snow(i)   = snow(i)   + snacc
               IF (snow(i).GT.snowmax(i)) THEN
                     snowmax(i)  = snow(i)
               ENDIF
               massbal(i)=massbal(i)+snacc
               enerbal(i)=sola(i)*(1.-albedo(i))+senflux(i)
     :                  +long(i)

C------------------------------------------------------------------
C  Refreezing of melt water
C     enerb1   =  energy availiable for melting and runoff
C     enerb2   =  heat flux into the ice
```

```
                  enerb1=enerbal(i)*exp(tice(i))
                  enerb2=enerbal(i)-enerb1
                  IF(enerbal(i).GT.0.) THEN
                        massbal(i)=massbal(i)-enerb1*.0026946*96
                  ENDIF
                  IF(enerbal(i).GT.0.) THEN
                        snow(i) = snow(i)-enerb1*.0026946*96
                  ENDIF
                  IF(snow(i).LT.0.) THEN
                        snow(i) = 0.
                  ENDIF

C-----------------------------------------------------------------------
C  Heating up 2 m of solid ice (density 900 kg/m^3)
C  Determining the albedo

                  IF(enerbal(i).GT.0.) THEN
                        tice(i) = tice(i)+enerb2*0.0002374*96
                  ENDIF
                  IF(tice(i).GT.0) THEN
                        tice(i) = 0.
                  ENDIF
                  albedo(i)=albsnow(i)-(albsnow(i)-albfin(i))*
     :                  exp(-snow(i)/200.)
                  IF(massbal(i).LT.0.) THEN
                        albedo(i)=albedo(i)+(massbal(i)
     :                        -snowmax(i))*0.000006
                  ENDIF
                  IF(albedo(i).LT.0.12) THEN
                        albedo(i)=0.12
                  ENDIF

                  precalt(i)  =  precalt(i) + precal
 1800       CONTINUE

C-----------------------------------------------------------------------
C  Print out a major block of output to a defined file every 10th day

            IF(klm.EQ.numrun)  THEN
                 IF(nday/10*10.EQ.nday) THEN
                    WRITE(50,2300) nday, (massbal(i), i=1,nx,2)
                    WRITE(50,2400) nday, (snow(i),    i=1,nx,2)
                    WRITE(50,2500) nday, (accum(i),   i=1,nx,2)
                    WRITE(50,2600) nday, (albedo(i),  i=1,nx,2)
                    WRITE(50,2700) nday, (tice(i),    i=1,nx,2)
                    WRITE(50,2750) nday, (precalt(i), i=1,nx,2)
                    WRITE(50, *)  ' '

                    WRITE(70,2755) nday, (enerbal(i), i=1,nx,2)
                    WRITE(70,2760) nday, (long(i),    i=1,nx,2)
                    WRITE(70,2765) nday, (senflux(i), i=1,nx,2)
                    WRITE(70,2770) nday, (latflux(i), i=1,nx,2)
                    WRITE(70, *)  ' '
                 ENDIF
            ENDIF
            nday =  nday + 1
            IF(nday.EQ.366) THEN
                 nday = 1
            ENDIF
            ndaytot =  ndaytot + 1
            IF(ndaytot.LE.ndaymax) THEN
                 GOTO 1700
            ENDIF
```

```
C---------------------------------------------------------------------------
C  Calculating mean specific balance, total precipitation and output of
C  the mass balance profile

          specbal = 0.
          prectot = 0.
          DO 1900 i = 1, nx
                specbal = specbal + area(i)*massbal(i)
                prectot = prectot + area(i)*precalt(i)
                j = nx + 1 - i
                WRITE(40, 2200) massbal(j),  alti(j)
 1900     CONTINUE

          specbal = specbal / totarea
          prectot = prectot / totarea

C---------------------------------------------------------------------------
C  Finding the equilibrium-line altitude (ella)

          DO 2000 i = 1, nxx
                a = massbal(i)
                b = massbal(i+1)
                IF(a.LT.0.AND.b.GE.0) THEN
                      ella =  alti(i) - (alti(i+1) -
     :                        alti(i))*a / (b - a)
                ENDIF
 2000     CONTINUE

          WRITE(40,2800) ella
          WRITE(40,2900) specbal
          WRITE(40,3000) prectot

          WRITE(*, *) 'Finished run ', klm
 2100 CONTINUE

 2200 FORMAT(1X, 2F10.0)
 2300 FORMAT('Mass balance', 1X, I4, 40F7.0)
 2400 FORMAT('Snow        ', 1X, I4, 40F7.0)
 2500 FORMAT('Accumulation', 1X, I4, 40F7.0)
 2600 FORMAT('Albedo      ', 1X, I4, 40F7.3)
 2700 FORMAT('Temp. ice   ', 1X, I4, 40F7.1)
 2750 FORMAT('Prec.       ', 1X, I4, 40F7.3)

 2755 FORMAT('Energy bal  ', I4, 40F7.1)
 2760 FORMAT('long        ', I4, 40F7.1)
 2765 FORMAT('senflux     ', I4, 40F7.1)
 2770 FORMAT('latflux     ', I4, 40F7.1)

 2800 FORMAT(1X, 'Equilibrium line at:', F7.1, 2X, 'metres')
 2900 FORMAT(1X, 'Mean specific balance:', F9.1, 2X, 'mm/yr')
 3000 FORMAT(1X, 'Mean precipitation:', F9.1, 2X, 'mm/yr' //)

      WRITE(*, 3100)

 3100 FORMAT(// 'FINISHED,  Please press return to end program.')

      PAUSE
      END
```

A1.2 The input files

Meteorological data input file
First line is:
Initial equilibrium altitude (section 5.4) Lapse rate Cloud base altitude Day start C_o C_1 C_2
Following lines:
Day Daily temperature Precipitation Relative humidity (fraction) Cloudiness (fraction)

For example:

```
721  -0.007  775   260    .950   ..
260     -4.10       .000    .690     .916
261      -.80       .000    .690     .666
262      1.90       .000    .805     .875
263      2.20       .000    .665     .666
264      1.10       .000    .745    1.000
265       .60       .100    .960    1.000
266      -.60      1.400    .990    1.000
                     .
                     .
                     .
253       .30       .000    .790    1.000
254      1.20       .000    .910     .957
255       .90      1.200    .990    1.000
256      1.50     10.200    .885     .500
257      -.90       .000    .700     .166
258      1.40       .000    .855    1.000
259      4.40       .100    .860    1.000
```

Glacier area distribution input file
First line is:
Total area number of segments
Following lines:
Area altitude of segment

For example:

```
6219       27
  26.   50
  66.   70
 116.   90
 122.  110
 157.  130
      .
      .
      .
 198.  450
 187.  470
 153.  490
 241.  525
 113.  575
  62.  625
  20.  675
```

Appendix Two
Meteorological data recorded at the Ny-Ålesund weather station for the seasons 1983 - 1988

This appendix presents some of the meteorological data used during this project. The plots are of the precipitation and daily average temperature recorded at the Ny-Ålesund weather station, north-west Spitsbergen, from mid-September to mid-September for the seasons 1983 to 1988.

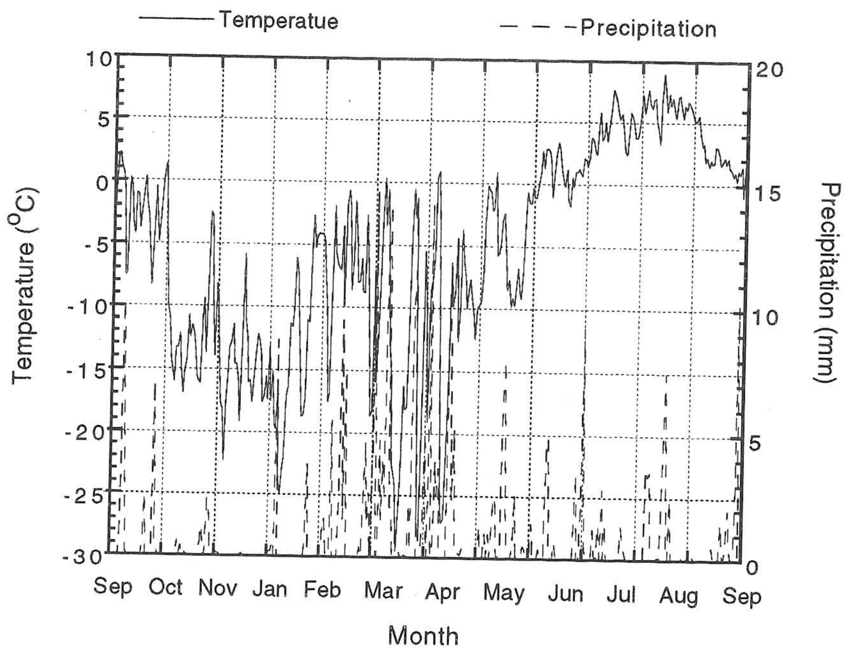

Figure A2.1 The temperature and precipitation for the season 1983/84 as recorded at the weather station at Ny-Ålesund.

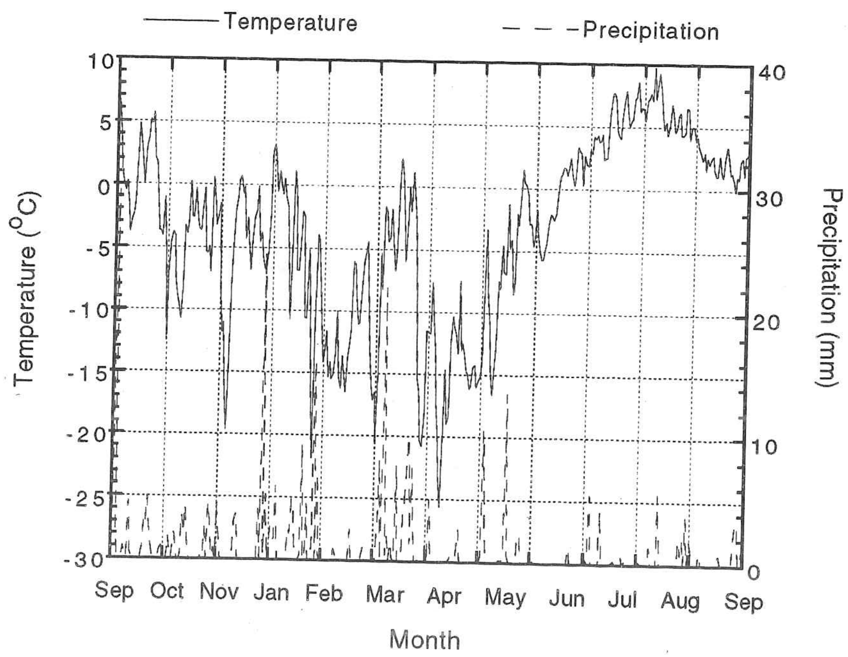

Figure A2.2 The temperature and precipitation for the season 1984/85 as recorded at the weather station at Ny-Ålesund.

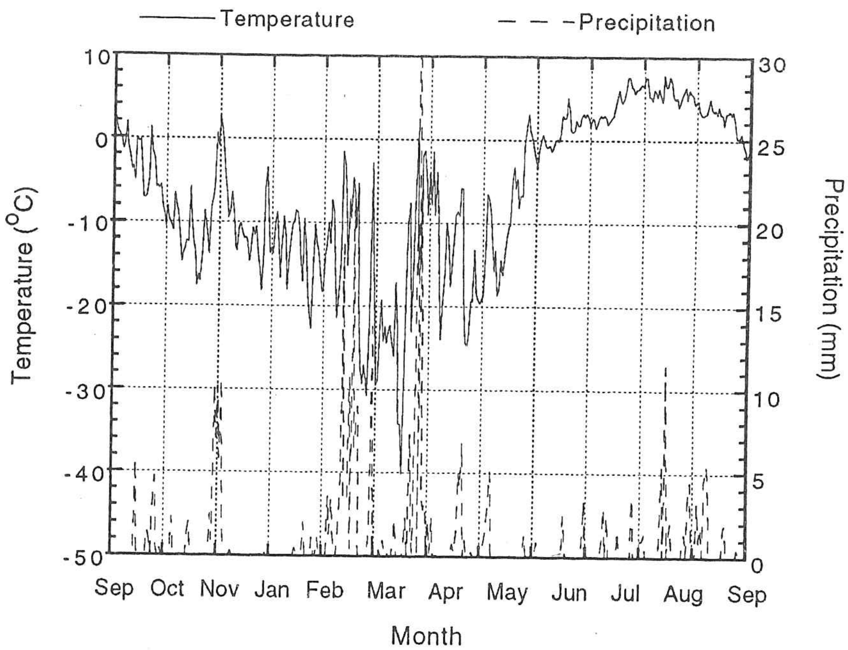

Figure A2.3. The temperature and precipitation for the season 1985/86 as recorded at the weather station at Ny-Ålesund.

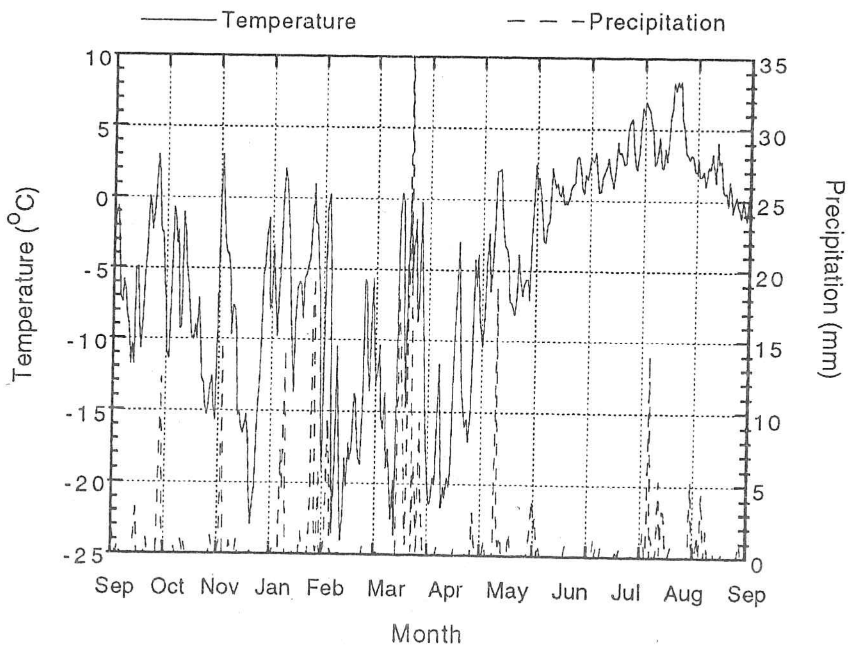

Figure A2.4 The temperature and precipitation for the season 1986/87 as recorded at the weather station at Ny-Ålesund.

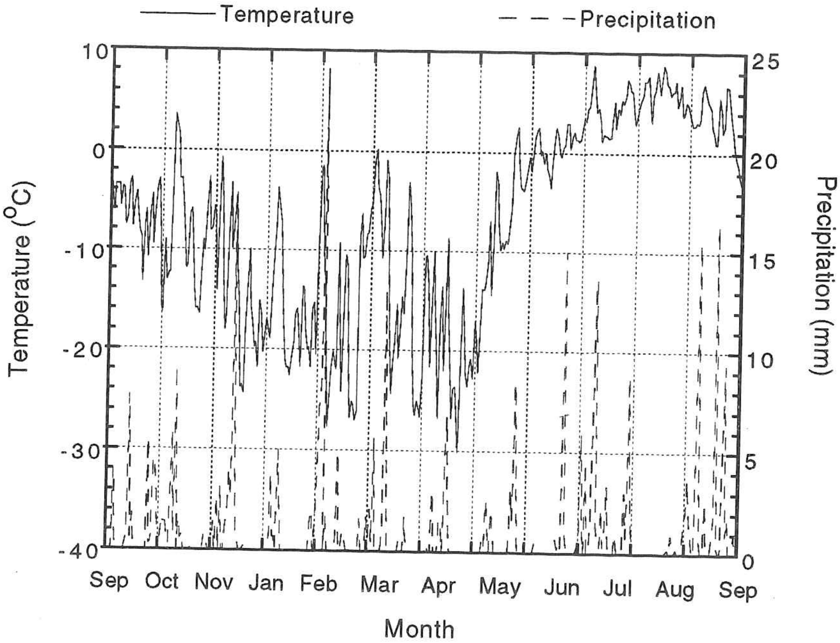

Figure A2.5 The temperature and precipitation for the season 1987/88 as recorded at the weather station at Ny-Ålesund.

Appendix Three
Sensitivity studies of Spitsbergen glaciers

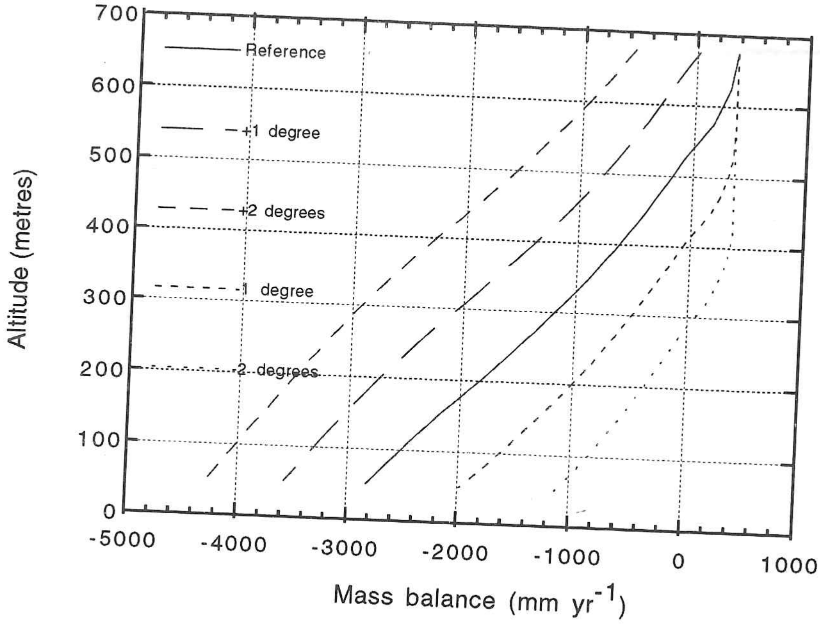

Figure A3.1 Mass balance profiles for changes in the annual mean temperature by +/- 1 to 2°C.

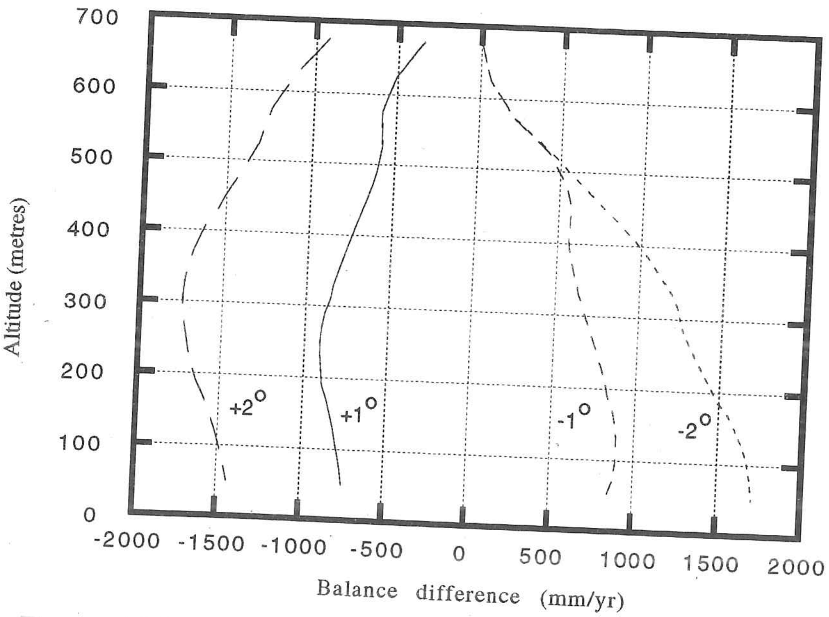

Figure A3.2 Variation in the specific mass-balance profiles for changes in the annual mean temperature by +/- 1 to 2°C.

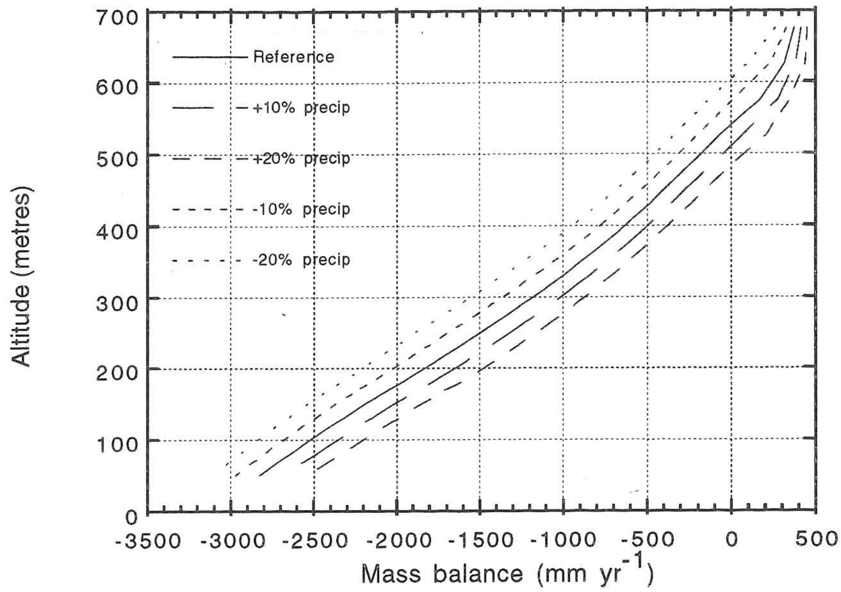

Figure A3.3 Mass balance profiles for changes in the annual precipitation by +/- 10 to 20%.

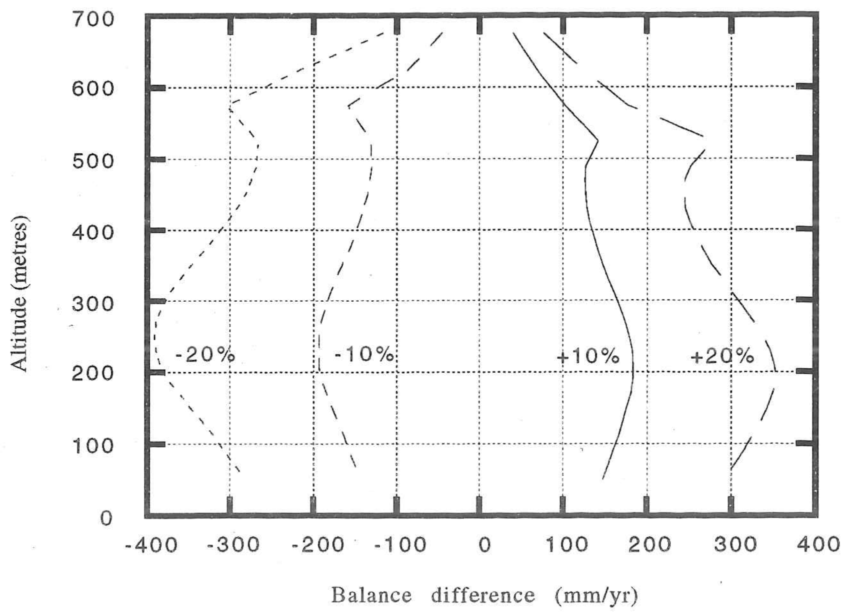

Figure A3.4 Variation in the specific mass-balance profiles for changes in the annual precipitation by +/- 10 to 20%.

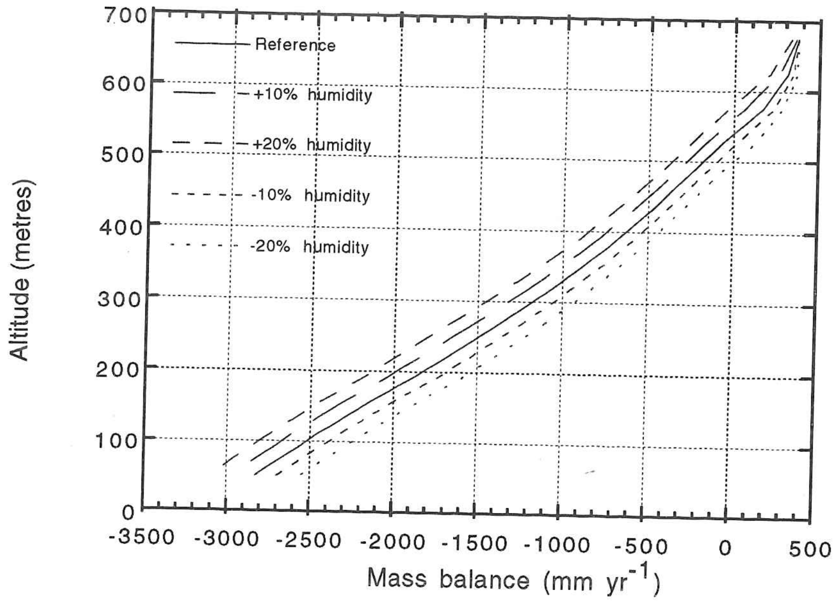

Figure A3.5 Mass balance profiles for changes in the annual average relative humidity by +/- 10 to 20%.

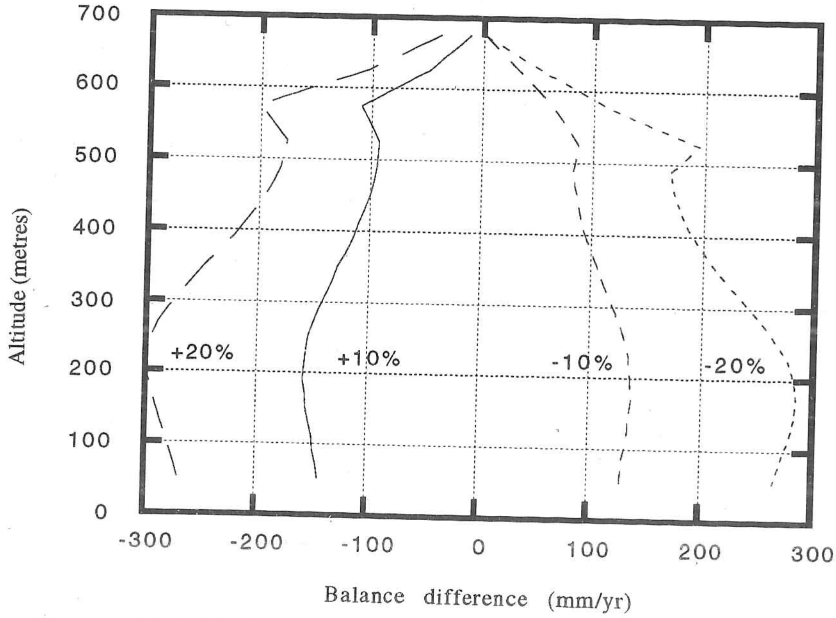

Figure A3.6 Variation in the specific mass-balance profiles for changes in the annual average relative humidity by +/- 10 to 20%.

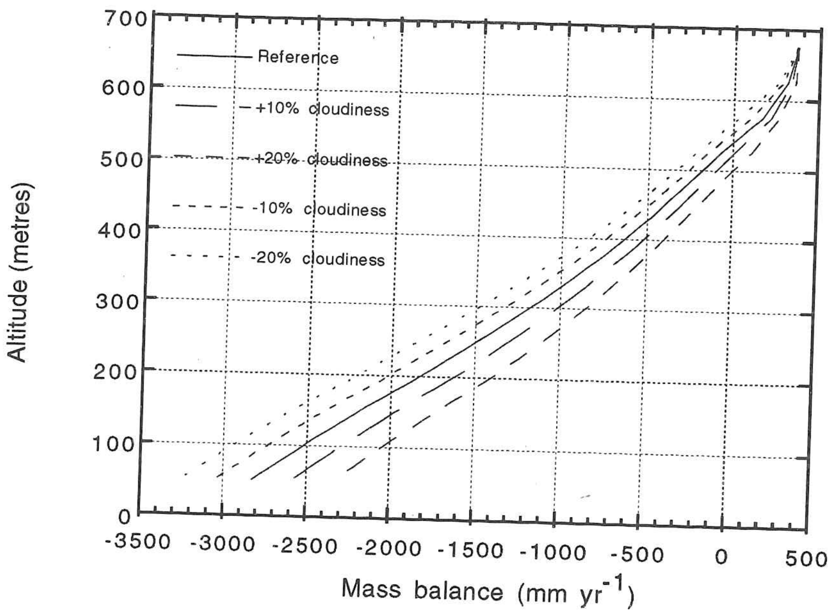

Figure A3.7 Mass balance profiles for changes in the annual average cloudiness by +/- 10 to 20%.

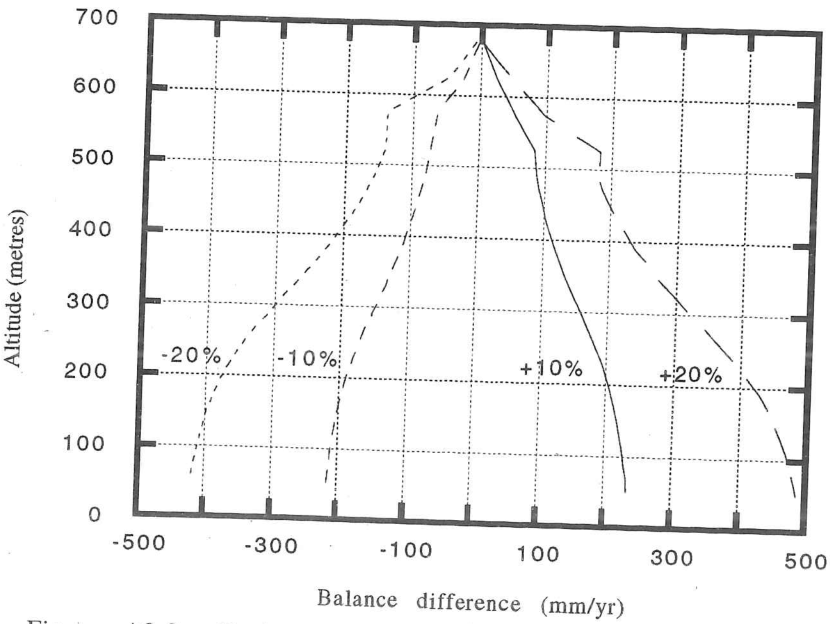

Figure A3.8 Variation in the specific mass-balance profiles for changes in the annual average cloudiness by +/- 10 to 20%.

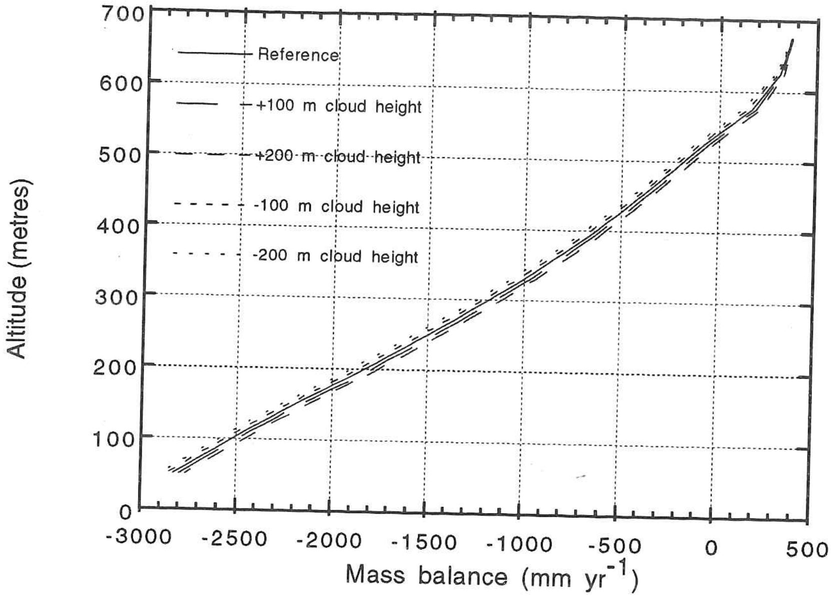

Figure A3.9 Mass balance profiles for changes in the height of the cloud base of +/- 100 to 200 metres.

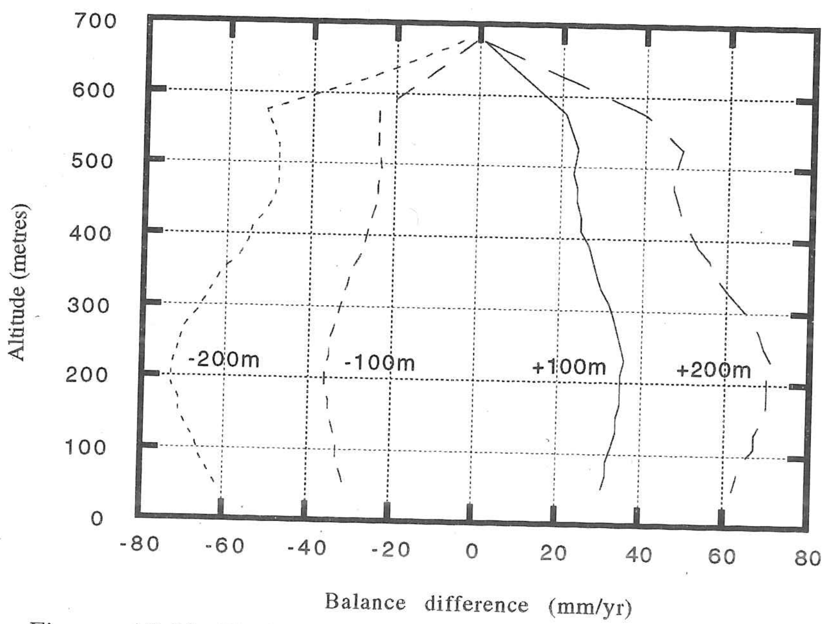

Figure A3.10 Variation in the specific mass-balance profiles for changes in the height of the cloud base of +/- 100 to 200 metres.

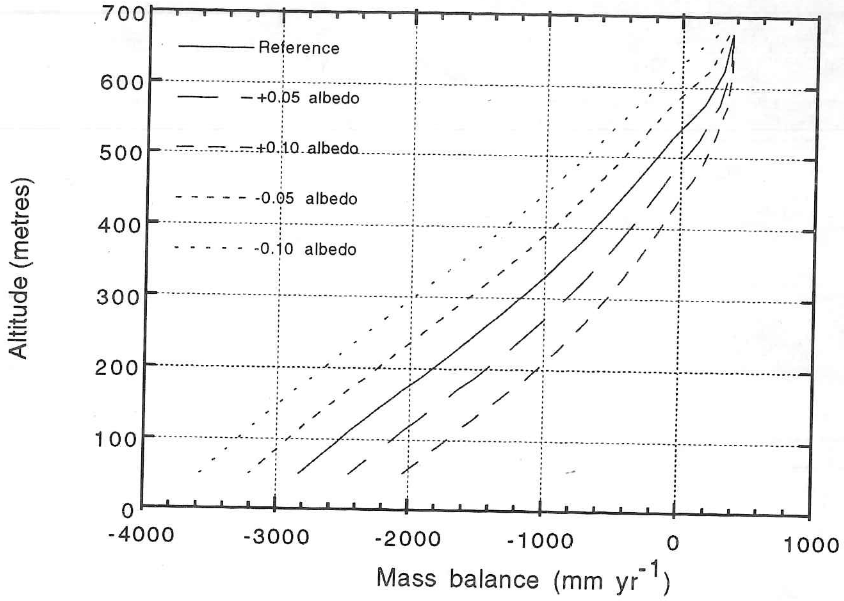

Figure A3.11 Mass balance profiles for changes in the albedo of the glacier by +/- 05 to 0.1.

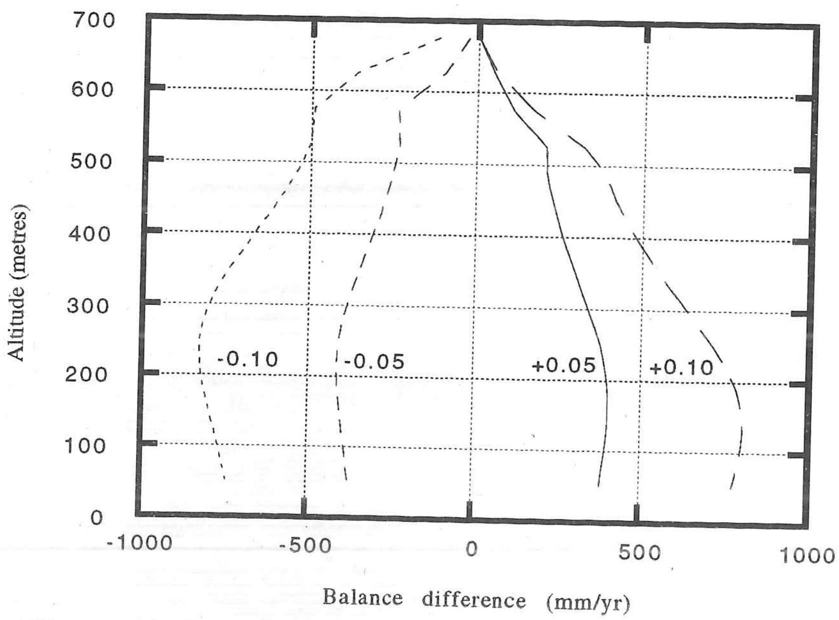

Figure A3.12 Variation in the specific mass-balance profiles for changes in the albedo of the glacier by +/- 05 to 0.1.